5 STRENGTH PATTERNS OF INTRAPLATE LITHOSPHERE

5.1 Introduction

Extensive geophysical work has revealed the heterogeneous thermal and compositional structure of continental lithosphere (e.g. Lithoprobe for western Canada [among others, Cook et al., 1992; Clowes et al., 1995] or for Europe [e.g. Blundell et al., 1992; Gee and Stephenson, 2006 and references therein]). The Canadian Cordilleran deep structure, for instance, has been successfully studied over the past decades with high quality geophysical probing. Seismic refraction and reflection studies [Kanasewich et al., 1994; Burianyk and Kanasewich, 1995; Cook, 1995; Clowes et al., 1995] have elucidated its crustal and lithospheric structure. In particular, considerable thinning of the lithosphere has been observed beneath the southern Canadian Cordillera associated with substantial thermal attenuation associated with elevated heat flow [Lewis et al., 1992; Hyndman et al., 1999]. The FFTB thus marks the transition zone between the disturbed (hot, thinned and deformed) lithosphere of the south-eastern Canadian Cordillera and the stable (cold and thick) North American Craton. The temperature structure and the mechanical response of the lithosphere to tectonic forces are interrelated with the rheological stratification forming the intermediate bridge.

Lithospheric rheological stratification depends firstly on its thermal and secondly on its layered compositional structure. Regional strength estimates are sensitive to large-scale temperature anomalies, especially of a hot mantle lithosphere [Ranalli, 1995; Burov and Diament, 1995]. Lithospheric strength may be inferred from the response of lithosphere to tectonic stresses and vertical loads, i.e. through cross spectral analyses of the gravity field [Flück et al., 2003; Pérez-Gussinyé and Watts, 2005]. Another approach is to calculate the mechanical stratification of the lithosphere in a forward manner as function of composition, pressures and temperatures and described by brittle and viscous rheological laws [e.g. Burov and Diament, 1995]. The mechanical stratification of the lithosphere, its response to tectonic or other loads as well as its thermal structure are tightly linked. This is most clearly evidenced by the different flexural response of hot versus cold lithosphere [e.g. Leever et al., 2006]. The south-eastern Canadian FFTB forms the supra-crustal deformation wedge that borders the Cordillera to the east. As such, the belt marks a transition between the westerly strongly deformed Cordillera interior and the stable continental platform of the North American Craton. This transition is not only noticeable in the supra-crustal deformation and denudation pattern (as the previous chapters addressed), but forms also a prime feature at lithosphere scale. The temperature evolution of the supra-crustal cover is tied to the lithosphere temperature state and basement heat flow pattern.

This study explores the 3-D lithospheric rheological structure to determine its spatial heterogeneity and allow for better comprehension of the observed deformation patterns in the FFTB and foreland basin to the east. The work entails a new work flow that was developed and tested on the west-central European lithosphere and afterwards applied in south-western Canada for modelling the thermal and rheological structure of the lithosphere [Cloetingh and Burov, 1996].
5 Strength patterns of intraplate lithosphere

5.2 Principles in lithospheric thermal and rheological structure

5.2.1 Lithosphere thermal structure

In this study the temperature structure of the lithosphere is calculated analytically taking surface heat flow and constraints of the thermal lithosphere base as boundary conditions. For continental lithosphere, this means a simple heat diffusion equation in which the surface heat flow \( q_{\text{surf}} \) serves as upper boundary condition. Furthermore, steady-state temperature conditions are assumed over a multilayer compositional model. The temperature distribution can consequently be estimated applying the Fourier’s law of heat conduction \[ \text{(5.1)} \]

\[
\frac{\partial q}{\partial t} = \kappa \frac{\partial^2 q}{\partial z^2} + A = 0,
\]

with \( q \) being the heat flow in W/m\(^2\), \( \kappa \) the thermal diffusivity in m\(^2\)/s and \( A \) being source term representing radiogenic heat production in W/m\(^3\). Heat transfer occurs in a 3-D

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Symbol</th>
<th>Units</th>
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<tbody>
<tr>
<td>Density</td>
<td>( \rho )</td>
<td>kg·m(^{-3})</td>
</tr>
<tr>
<td>Thermal conductivity</td>
<td>( k )</td>
<td>W·m(^{-1})·K(^{-1})</td>
</tr>
<tr>
<td>Specific heat</td>
<td>( c_p )</td>
<td>J·kg(^{-1})·K(^{-1})</td>
</tr>
<tr>
<td>Heat production</td>
<td>( H )</td>
<td>W·m(^{-3})</td>
</tr>
<tr>
<td>Exp. decay rate of heat prod. (skind)</td>
<td>( z_H )</td>
<td>km</td>
</tr>
<tr>
<td>Power law exponent</td>
<td>( n )</td>
<td>-</td>
</tr>
<tr>
<td>Power law activation energy</td>
<td>( E_p )</td>
<td>kJ·mole(^{-1})</td>
</tr>
<tr>
<td>Power law strain rate</td>
<td>( A_p )</td>
<td>Pa(^n)·s(^{-1})</td>
</tr>
<tr>
<td>Dorn law activation energy</td>
<td>( E_D )</td>
<td>kJ·mole(^{-1})</td>
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<tr>
<td>Dorn law strain rate</td>
<td>( A_D )</td>
<td>s(^{-1})</td>
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<td>Dorn law stress</td>
<td>( E_D )</td>
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<table>
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<tr>
<th>Parameter</th>
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<tbody>
<tr>
<td>Sediment</td>
<td>Upper crust</td>
</tr>
<tr>
<td>Various</td>
<td>2400</td>
</tr>
</tbody>
</table>

Thermal and rheological model parameters for European strength-map (case I) calculations. Power law creep parameters \((n, E_p, A_p)\) for quartzite upper crust, and diorite/diabase lower crust from Carter and Tsenn [1987]. The Power and Dorn law creep parameters for the Olivine upper mantle come from Goetze and Evans [1979].
compositionally and thermally variable lithosphere but is commonly reduced to a 1-D problem considering that vertical heat flow dominates. A temperature structure as function of depth follows that can be given by the following analytical expression [e.g. Cermak and Bodri, 1995]:

\[ T(z) = T_{surf} + \frac{q_{surf}}{k} - \frac{A}{2k} z^2, \quad (5.2) \]

with \( T(z) \) being the temperature as function of depth, \( T_{surf} \) the temperature at surface, \( k \) the thermal conductivity in W·m\(^{-1}\)·K\(^{-1}\), and \( A \) the heat production by radioactive decay in W/m\(^3\). These thermal rock parameters are strongly dependant on the lithologies of which some general estimates are given in Table 5.1. Shallow surface heat generation estimates are commonly extrapolated assuming a linear or exponential decrease with depth [Parson and Sclater, 1977; Turcotte and Schubert, 2002].

Alternative to simple 1-D steady-state temperature conditions, non-steady state or multi-dimensional temperature problems may be approximated using a finite-difference approach. For oceanic lithosphere or strongly thermally perturbed continental lithosphere a forward cooling model can be adopted [Parson and Sclater, 1977]. The thermal lithosphere can be understood as a thermal boundary layer of an infinite half-space with a fixed temperature at the upper boundary interface where heat is lost. From an initially hot and thin thermal boundary layer the system cools and thickens downward. The temperature structure of oceanic or thermally perturbed continental lithosphere can be characterized by the thermal age since the last heating event occurred and cooling started.

The base of the thermal lithosphere is commonly identified with an isotherm, which characterizes the thickness of the conductive boundary layer [Turcotte and Schubert, 2002]. An isotherm at 1300 °C is typically considered separating the thermal lithosphere where heat transfer primarily occurs by conduction from the asthenosphere where convection prevails [Schmeling and Marquart, 1993].

5.2.2 Rheological laws

Strength can be defined as the resistance to deformation, limited by the differential stress required to initiate failure. This is most clear under brittle conditions, where failure occurs at the yield stress. For viscous deformation instead, flow is not limited to a yield stress, although a level of stress is required to allow flow at a predefined threshold strain rate. Rock strength may be defined by the deformation mechanism for which, under given conditions, the least amount of differential stress is required. At low temperatures, equivalent to shallow depths, rocks predominantly deform by brittle mechanisms, i.e. fracture and frictional sliding. Brittle deformation is strain rate and temperature independent and generally described by Byerlee’s law, defined for frictional sliding on pre-existing fractures [Byerlee, 1978; Ranalli and Murphy, 1987; Kohlstedt et al., 1995; Ranalli, 1995]. This can be written in terms of the principal stress difference (\( \Delta \sigma = \sigma_1 - \sigma_3 \)) under which slip on pre-existing weakness zones begins:

\[ (\sigma_1 - \sigma_3) = \alpha \cdot \rho \cdot z \cdot (1 - \lambda), \quad (5.3) \]

with \( \alpha \) being the tectonic regime coefficient (taken as 1.67 and 0.67 for compressional and tensile regimes, respectively), \( \lambda \) is the pore fluid factor, and \( \rho \cdot g \cdot z \) is the overburden pressure. It should be noted that the brittle strength under compression is a factor three larger than the tensile strength.
At greater depths, i.e. at higher temperatures and confining pressure, creep mechanisms prevail, which are described by ductile flow laws. These rheological laws are based on both experimental rock mechanics and modelling studies [Kirby and Kronenberg, 1987; Carter and Tsenn, 1987; Wilks and Carter, 1990; Kohlstedt et al., 1995; Ranalli, 1995]. The predominant creep mechanism in the ductile parts of the lithosphere and at deformation rates typical for plate tectonic deformation ($10^{-15}$ to $10^{-17}$ s$^{-1}$) can be described by a power law constitutive equation:

$$\sigma_i - \sigma_j = (\dot{\varepsilon} \cdot A_p)^{1/n} \cdot e^{[E_p/nRT]}$$

with $(\sigma_i - \sigma_j)$ the differential stress (under both compression and extension), $\dot{\varepsilon}$ the strain rate, $T$ the temperature, $R$ the universal gas constant, and $A_p$, $E_p$ and $n$ material dependent creep parameters (see Table 5.1).

For olivine-dominated subcrustal lithosphere power law creep is the preferred deformation mechanism for differential stresses up to 200 MPa. For larger stresses, creep deformation is better described by a Dorn law constitutive equation:

$$\sigma_i - \sigma_j = \sigma_0 \cdot [1 - (RT/E_0) \cdot \ln (\dot{\varepsilon}/A_0)]^{1/2}$$

where $E_0$, $A_0$ are material dependent Dorn creep parameters (see Table 5.1).

The strength of the lithosphere varies considerably with depth as a function of confining pressure, temperature and composition. The presence or absence of fluids and rate of deformation also play an important role [e.g. Carter and Tsenn, 1987; Kirby and Kronenberg, 1987].

### 5.2.3 Vertical rheological structure of lithosphere

The strength of continental lithosphere is controlled by its depth-dependent rheological structure in which the thickness and composition of crustal layers, crustal to upper mantle temperature distribution and strain-rates play dominant roles [e.g. Carter and Tsenn, 1987; Kirby and Kronenberg, 1987] (Figure 5.1a). The rheological stratification of the multi-layer continental lithosphere differs widely from the ocean lithosphere, the later comprising a single layer with olivine rheology with a brittle and ductile branch. Instead, a typical rheological model of a multi-layer continental lithosphere (Figure 5.1a) consists of a mechanically strong upper crust, which is separated by a weak lower crustal layer from the strong upper part of the mantle lithosphere. The lower crust is predominantly in a ductile regime as the creep activation of its main constituents (i.e. quartz or feldspar) occurs at temperatures of 400-600 °C [Toussaint et al., 2004]. The mechanically strong mantle lithosphere in turn overlies a weak lower mantle lithosphere [Cloetingh and Burov, 1996].

The product of a multilayer quartzite (upper crust), diorite (lower crust) and olivine (mantle) rheologies are reflected in the brittle and ductile branches of the strength envelopes that follow the Byerlee failure criteria and the Power law and Dorn law ductile creep (Figure 5.1a). The change from brittle to ductile dominated deformation is referred to as the brittle-ductile transition. Furthermore, mechanical strength may be defined as strength in excess of 10-20 MPa [Burov and Diament, 1995] with mechanically strong layers having thicknesses $h_{uc}$, $h_{lc}$ and $h_{ml}$ for the upper, lower crust and mantle lithosphere, respectively (Figure 5.1a).

The strength distribution over depth for different types of lithosphere is controlled by the compositional layering and temperature structure of the lithosphere (Figure 5.1b).
Thick cratonic lithosphere exhibits cold crust and mantle lithosphere. Consequently, brittle strength dominates to greater depths and ductile creep only occurs at greater depth where temperatures are higher. In contrast, strength in hot backarc or destabilized orogenic lithosphere, where the mantle lithosphere is severely thinned, resides mainly in the upper crust. The time lapse since the thermal instability occurred is an important parameter for these hot continental realms. Thermal stabilization of continental lithosphere occurs over time spans of 150-200 Ma before reaching thermally and rheologically established lithosphere [e.g., Cloetingh and Burov, 1996]. Oceanic lithosphere is compositionally more homogeneous, its olivine-dominated rheology depending on its age-dependent temperature structure [Panza et al., 1980; Stephenson and Cloetingh, 1991; Cloetingh and Burov, 1996].

5.2.4 Mechanically strong layering and EET estimates

When subject to horizontal or vertical forces (e.g. topographic loads or tectonic stresses), the lithosphere exhibits elastic behaviour in the form of flexure. The flexural response to loading provides information on the mechanical properties of the lithosphere. For instance, inferences can be made on the thickness and coupling of mechanically strong layering with the study of the flexural response of the lithosphere. In particular, regional

![Conceptual chart of rheological structure](image)

Conceptual chart of rheological structure, depicting consequences of variations in rheological parameters for the mechanical strength of the lithosphere. (a) Schematic representation of the lithospheric strength structure and mechanically strong mantle lithosphere as function of a conductive temperature profile for a multilayer lithosphere. (b) Strength envelopes for (I) a single-layer oceanic lithosphere, (II) a thin and hot continental backarc lithosphere and (III) a cold cratonic lithosphere.
Strength patterns of intraplate lithosphere subsidence and uplift trends exhibit the elastic behaviour of the lithosphere in the form of foreland basins or flexural rebound effects. The foreland basin development of the south-eastern Canadian Cordilleran, for instance, entails information on the mechanical behaviour of its lithosphere [Beaumont, 1981; Stockmal et al., 1992; Peper, 1993]. Also the large-wavelength post-orogenic (post-Paleocene) uplift discussed in Chapter 4 may be attributed to elastic response as rebound to unloading of the Cordillera interior. While large-wavelength vertical motions and topographic features constitute noticeable features which are well explained by the elasticity of the lithosphere, gravity field studies provide often better means to examine the mechanical structure from the lithosphere response [e.g. Stephenson and Lambeck, 1985; Burov et al., 1993; Flück et al., 2003, Pérez-Gussinye and Watts, 2005].

The response function of the lithosphere thus provides estimates of the effective elastic thickness, defined as a thickness of an equivalent elastic plate [Watts, 1978, 1992, 2001]. The effective elastic thickness (EET) may be translated to the combined layer thickness of the involved mechanically strong layers [Burov and Diament, 1995; Watts and Burov, 2003]. Effective elastic thicknesses may not only be estimated from foreland basin deflection studies but also derived from calculating mechanically strong layer thicknesses.

Layers within the lithosphere that can support stresses in excess of 10-20 MPa [Cloetingh and Burov, 1996; Burov and Diament, 1995], have been loosely defined as mechanically strong layers. If the strong crustal and mantle layers are mechanically coupled, the continental lithosphere will deform as a single plate. For a coupled rheology, the mechanically strong crust and mantle are ‘welded’ and the EET can be estimated from

\[ T_e \approx h_{UC} + h_{LC} + h_{ML}, \]  

where \( h_{UC}, h_{LC} \) and \( h_{ML} \) are the thicknesses of the mechanically strong layers (Figure 5.2b) for the upper, lower crust and mantle lithosphere, respectively. On the other hand, decoupling of the different mechanically strong layers, especially between the crust and mantle, drastically reduces (by around two times) the EET [Burov and Diament, 1995]. The EET for three detached layers is given by

\[ T_e \approx \sqrt[3]{h_{UC}^3 + h_{LC}^3 + h_{ML}^3}. \]  

Furthermore, effective elastic thicknesses may be estimated from coherence and admittance modelling of gravity anomalies to topography spectral trends [Stephenson and Lambeck, 1985; Burov et al., 1993; Flück et al., 2003; Pérez-Gussinyé and Watts, 2005]. The flexural rigidity \( D \) defines the resistance to bending of an elastic plate with a laterally uniform elastic thickness as given by:

\[ D = \frac{E T_e^3}{12(1 - \nu^2)}, \]  

in which the elastic parameters \( E \) and \( \nu \) are respectively the Young’s modulus and Poisson’s ratio [e.g. Turcotte and Schubert, 2002]. The wavelength of the down-flexure of a simple 2-D elastic plate under a point load is given by:

\[ \lambda = 2\pi \left( \frac{D}{g\Delta \rho} \right), \]  

in which \( g \) is gravitation constant and \( \Delta \rho \) the density contrast between the materials below and above the elastic plate [Turcotte and Schubert, 2002].
5.3 Modelling work-flow for 3-D strength model

The workflow for lithospheric strength-map compilation has been first built and tested on the European lithosphere [Hardebol et al., 2003; Cloetingh et al., 2005, 2006; Tesouro et al., 2009]. This study aims at quantifying the 3-D spatial variability of lithospheric strength, starting from rheological laws of brittle and ductile deformation [Kohlstedt et al., 1995; Ranalli, 1995; Kirby and Kronenberg, 1987; Wilks and Carter, 1990; Carter and Tsenn, 1987]. The workflow integrates a large body of geophysical and geological data with the aim to produce a constrained 3-D strength model to allow for comparison with the first order tectonic and strain patterns.

First, compositional and thermal models are compiled and used jointly as input for the subsequent construction of a 3-D strength model (Figure 5.2). The 3-D temperature and strength models are compiled by calculating 1-D depth profiles for each grid point.
covering the study area (Figure 5.2). Calculations are done using the Geographic Resources Analysis Support System (GRASS), which is a Geographic Information System (GIS) that facilitates retrieval of 2D and 3-D input files from a spatial database that handles projection and raster resolution aspects. Integrated strength maps are derived from the 3-D strength model by integrating over its full depth range. In addition, normalized strength maps can be calculated that depict the strength for the different compositional layers separately.

### 5.3.1 Compositional input model

The 3-D compositional model may comprise continental lithospheric domains of various types as well as oceanic lithosphere. For continental lithosphere a multi-layer model is adopted with the crust being divided into upper and lower crust of equal thicknesses, overlain by sedimentary cover for places where it reaches a sufficient thickness.

The model comprises lateral uniform compositional layers. A quartzite composition is adopted for the upper crust, a diorite composition for the lower crust, with a lithospheric mantle that essentially consists of dry olivine. The oceanic lithosphere comprises the single layer of olivine mantle lithosphere. The mantle of both continental and oceanic lithospheres exhibits a uniform olivine composition down to the base of the model domain at 250 km. The base of the mantle lithosphere is a thermal boundary that results from the thermal modelling and is not set by a compositional change. For the thermal and rheological modelling, rock parameters are added to the compositional layers as listed in Table 5.1 and taken from Cloetingh and Burov [1996].

As mentioned earlier, the lithospheric strength calculation (with the ductile component set by a threshold strain rate) is primarily determined by the temperature structure of the mantle and secondly by the composition and thickness variations of the crust. The compositional input models in this study yield a layer-cake with uniform composition. The effect of compositional heterogeneity of the crust for this study is mainly due to thickness variations where weak lower crust is replaced by much stronger olivine mantle material. In addition, the calculated strength also alters as the temperature input model is affected by changes in the conductive heat transfer and heat generation from varying compositional layer thicknesses.

### 5.3.2 Thermal input model

The conductive heat equation (Equation 5.1) is solved with surface temperature and heat-flow as boundary conditions and adding thermal conductivity parameters to the constituent compositional layers (Table 5.1). An exponentially decreasing heat production is adopted for the sedimentary and crustal layers \((A_z = A_0 \exp(-z/D))\) where \(A_z\) is heat generation as function of depth, \(A_0\) the heat generation at layers top and \(D\) the depth scaling factor [Turcotte and Schubert, 2002]. Calculated lithosphere temperature depth profiles are evaluated against alternative constraints on the thickness of the thermal lithosphere. The construction of the 3-D thermal model consequently involves an iterative process comparing the analytical solution set by the surface heat flow boundary condition and alternative constraints on the deeper thermal structure.

Temperatures can only be directly measured in boreholes limited to the few upper kilometres. Raw borehole temperatures are affected by paleoclimate and conductivity heterogeneities, for which different corrections can be applied [e.g. Pollack et al., 1993]. Furthermore heat flow estimates may contain artefacts due to non-steady state or non-
conductive conditions from for instance fluid flow or strong denudation or sedimentation effects. Recent heat flow compilations have undergone much correction, whereas values with strong local noise have been filtered from regional or global databases.

The deep temperature structure of the mantle lithosphere can only be probed with indirect proxies like seismic velocities. Like with other geophysical proxies (e.g. magnetotelluric sounding [e.g. Majorowicz and Gough, 1991; Hjelt and Korja, 1993], strong dependency on temperature is essential [Bijwaard and Spakman, 1999, 2000; Goes et al., 2000a, 2000b]. It is widely acknowledged that composition, presence of partial melts or water and anisotropy also impact seismic velocities substantially [Goes et al., 2000a; Artemieva et al., 2006]. At least for the crust, compositional heterogeneities exceed temperature effects on seismic velocities.

A map of the calculated base of the thermal lithosphere provides a measure of the spatial variations in the thermal model. It also allows comparison with the other proxies for the base of the thermal lithosphere. The base of the seismic lithosphere is often associated with the top of the low velocity or viscosity zone (LVZ) and with temperatures near the mantle solidus associated with partial melts. Thereby, one can argue for an intimate linkage between the thermal base of the lithosphere and seismic or alternatively electromagnetic observations [Schmeling and Marquart, 1993, Artemieva and Mooney, 2001; Plomerova et al., 2002] and thus constraining the depth of the 1300 °C isotherm [see also Artemieva et al., 2006].

5.3.3 Restrictions of the thermal and rheological modelling

The 3-D modelling comprises extended regional scales aiming at first-order appraisal of lateral strength discontinuities. For this reason the 3-D temperature model does not include the second order effects of lateral heat conduction that especially affects regions of strong lateral temperature contrasts. In addition, simplified steady-state conditions were assumed, omitting transient phenomena associated with mantle plumes or orogenic areas with thick thermally unstable crustal roots, or areas of fast sedimentary burial or erosion. Under these circumstances, surface heat flow alone is an inadequate proxy for the characterization of the thermal state of the lithosphere, in particular at upper mantle depths. As pointed out above, the thermal structure for these areas is calibrated using seismological constraints to derive a more precise temperature structure. Uncertainties in the thermal model inevitably enter into the strength modelling and are further amplified by uncertainties in the compositional model.

The bulk strength that is reflected in the long-term and first-order deformation patterns is an approximate description of the complex stratified ductile and viscous lithosphere. As the stress fields of intraplate lithosphere are largely compressional [Zoback, 1992], regional estimates of brittle strength are given for the compressive mode. Nonetheless, strength envelopes, while representing a simplification of the actual rheological state and response, have shown being useful for describing the bulk mechanical properties of the lithosphere. Recent discussions about the presence and meaning of a mechanically strong mantle and the relationship between seismicity, strength and deformation mechanisms and the character of brittle-ductile transitions have highlight the shortcomings and merits of strength envelopes [Jackson, 2002; Handy and Brun, 2004; Burov and Watts, 2006].
Compositional and thermal input map for the European strength-map compilation used in the workflow outlined in Figure 5.2. (a) Compositional input map depicting depth to Moho (Dmoho) for definition of base of lower crust (DMoho) and base upper crust (DMoho/2). The greyish overlay shows oceanic lithosphere for which a 1-layer model is adopted. (b) Thermal input map. For the continental domain surface heat flow values (mW/m²) are used for calculation of steady-state geotherms. For the oceanic domain, ages of the last thermo-tectonic event are adopted when applying a cooling model (grey-scale). (c) The 1300 °C isotherm map from the 3-D thermal model that serves as input for strength calculations. The grey contouring depicts alternative estimates of the depth to the base of the lithosphere [Babuska and Plomerova, 1992, 2001].
5.4 Case 1: rheological structure of European lithosphere

Numerous studies have calculated strength profiles and lithospheric effective elastic thicknesses for many locations in the heterogeneous European lithosphere [e.g. Cloetingh and Burov, 1996]. Most of these locations are along available deep seismic crustal cross sections, such as the European Geotraverse [Cloetingh and Banda, 1992] and the TransAlp deep seismic profile [Willingshofer and Cloetingh, 2003]. Previous lithospheric strength maps have been calculated only for regional areas of Europe, including the Pannonian Basin-Carpathian region [Lankreijer et al., 1999] and the Baltic Shield [Kaikkonen et al., 2000; Moisio et al., 2000; Artemieva and Mooney, 2001].

More recently, the rheological structure of the European lithosphere is examined in a more integral manner [see especially the work of Tesauro et al., 2009], complying with the workflow and exploratory strength-map compilation of this study [Hardebol et al., 2003; Cloetingh et al., 2005, 2006]. The 3-D rheological modelling results are presented as integrated strength-maps displaying well the spatial variations across Europe in correlation with the major structural tectonic features. A regularized grid of 204 by 199 cells is applied in the easting and northing direction, respectively, giving resolution of 15 km per cell. The study area is projected using a Lambert Azimuthal Equal Area projection, centred at 4°E, 48°N and with a lower left corner at 33°N, 20°W and upper right corner at 62°N, 36.5°E.

Drawing on the newly compiled Moho map of Dèzes and Ziegler [2004], constraints on the thermal lithospheric structure from heat flow studies, and estimates of lithospheric thickness from seismological studies [Plomerova et al., 2002], a 3-D strength model of the lithosphere of Europe is constructed (Figure 5.4). The rheological strength model is based on a compositional model including one upper mantle layer and two crustal layers [Hardebol et al., 2003].

5.4.1 Input models

The European Moho map of Dèzes and Ziegler [2004] (Figure 5.3a) was used as main resource for the definition of the crustal layers. The map is constructed from a range of data sources including several regional and European scale compilations of deep seismic reflection, refraction and surface wave dispersion studies [e.g. Panza et al., 1980; Blundell et al. 1992]. The Moho map is extended eastward to include the Carpathians and Eastern European Craton [Lenkey, 1999].

The study also used regional scale compilations of seismologically defined lithosphere thickness, which correlate with the thickness of a thermal boundary layer. Comparison with the calculated base of the thermal lithosphere gives generally good agreement (see discussion on thermal input models below) [Panza et al., 1980; Babuska and Plomerova, 1993, 2001; Artemieva et al., 2006].

For the thermal input model, surface heat flow constraints (Figure 5.2b) were taken from the Global Heat Flow Data Set [Pollack et al., 1993], the Geothermal Atlas of Europe [Hurtig et al., 1992] and several regional surface heat flow studies from France [Lucazeau and Vasseur, 1989], the Alpine-Carpathian system including the Pannonian Basin [Lenkey, 1999; Tari et al., 1999; Sachsenhofer, 2001], southwest Poland [Majorowicz, 2004] and the Iberian Peninsula [Fernández et al., 1998; Majorowicz, 2004]. For the oceanic realm an input raster is provided (Figure 5.2b. – in grey scale), outlining the thermal age with Neogene ages for the Mediterranean region, e.g. Valencia Trough, and Jurassic to Cretaceous ages for the Atlantic Ocean and Rockall Trough.
5 Strength Patterns of Intraplate Lithosphere

Beside these temperature input data for resolving the heat equation, also proxies of the deeper temperature structure of the lithosphere are considered. For instance, the depth to the thermal base of the European lithosphere can be derived from surface wave dispersion or seismic anisotropy studies [Babuska and Plomerova, 1992, 1993] and Plomerova et al. [2002], and from electrical or magnetotelluric sounding studies [e.g. Praus et al., 1990; Hjelt and Korja, 1993; Adam and Wesztergom, 2001]. Also, the inversion of tomographic models into temperatures provides depth slices of temperatures at 52.5, 95, 145 and 200 km [Goes et al., 2000a, 2000b]. The calculated temperature model is being compared against these proxies of the deeper lithosphere temperature structure. The model is updated through a few iterative steps for a few areas where the steady state and surface boundary condition clearly misrepresent the deeper lithosphere structure. This is done to account for the fact that the heat equation is essentially solved with the surface temperature and surface heat flow as boundary conditions, and unless through an iterative update, indirect proxies from the base of the lithosphere are otherwise not weighted.

5.4.2 Thermal model description

Seismic tomography indicates that beneath the Massif Central, the thermal thickness of the lithosphere is reduced to 50-60 km [Sobolev et al., 1997], increasing both eastward to about 110 km beneath the Western Alps [Lippitsch et al., 2003] and westward to 120 km beneath coastal France [Dèzes et al., 2004]. Northward, seismic constraints on the thermal thickness indicate 120-150 km under the Paris Basin, decreasing to 50-60 km beneath the Rhenish Massif [Babuska and Plomerova, 1992; Prodehl et al., 1995; Goes et al., 2000a, 2000b] and reaching again about 80 km beneath the Eger Graben [Babuska and Plomerova, 2001]. Beneath the Upper Rhine Graben, the lithosphere reaches a thickness of about 100 km [Achauer and Masson, 2002] and increases eastward to about 120 km in the area of the Franconian Platform [Babuska and Plomerova, 1992].

Comparison of the depth distribution to the 1300 °C isotherm from the 3-D temperature model with seismic constraints (Figure 5.2c) shows a good first order resemblance. Also for the Fennoscandian Shield, the Eastern European Craton and Paleozoic inliers such as the Bohemian Massif, London-Brabant Massif and Armorican Massif with stabilized upper mantle temperatures, our model correlates well with low temperatures and deep position of the thermal base of the lithosphere from tomography models.

For a number of regions however, our compiled surface heat flow input model does not perfectly represent the thermal perturbation inferred from mantle tomography [Goes et al., 2000a, 2000b]. For example, a poorly constrained thermal anomaly underneath Southern Norway [Rohrman et al., 2002] has no equivalent in our model. Furthermore in areas of the European Cenozoic Rift System, the effect of strong fault controlled fluid flow or thermal blanketing through the accumulation of thick syn- and post-rift sedimentary sequences can alter the crustal temperature structure. Consequently, surface heat flow values no longer reflect a wider lithosphere scale thermal structure. For example, in the Rhine Graben region, the 1300 °C isotherm representing the base of the thermal lithosphere is calculated to correspond to a depth of approximately 65-70 km, whereas from seismological studies a depth of 80 to 90 km is inferred. Because hot fluids circulating through crustal scale faults elevated the surface heat flow estimates, the actual temperature in the upper mantle tend to be more moderate than predicted from our steady-state calculations.
Nevertheless, the followed approach is considered reliable enough in the scope of a first order continent-wide appraisal. A forward modelled temperature model tested and iteratively corrected to alternative and independent surface and deeper remote sensing data circumvents the limitations of the individual data-sets as proxy of the lithospheric thermal structure. In follow-up studies, the iterative calibration of strong thermally perturbed regions may adopt non-steady-state conditions.

5.4.3 Results from the rheological reference model

A rheological reference model (Figure 5.4) is calculated that is based on the first order features in compositional and thermal structure. A constant bulk strain rate is adopted, like in previous regional rheological models for which different values in the range of $10^{-16}$ to $10^{-14}$ s$^{-1}$ have been used [e.g. Carter and Tsenn, 1987; Lankreijer et al., 1999; van Wees and Beekman, 2000]. For the here presented rheological model a strain rate of $10^{-15}$ s$^{-1}$ is.

The depth-varying rheology employed in this study depends on several variables, of which the most important are crustal thickness, composition and temperature. The creep parameters for the different compositional layers are given in Table 5.1.

Figure 5.4 shows the calculated integrated compressional strength of the European lithosphere. The lithosphere shows major spatial strength variability, with a pronounced contrast between the north-eastern cratonic regions of high rigidity northeast of the Teisseyre-Tornquist Zone and the regions to the southwest. The latter shows a patchwork pattern of lithospheric strength variability that strongly reflects its polyphase Phanerozoic history. It comprises patches of thermally stabilized, formerly accreted Variscan terranes, enclosed by partly recalibrated Mesozoic extensional structures and overprinted by recent Cenozoic thermally perturbed weak zones.

The highest integrated strengths are obtained for the Fennoscandian Shield and adjacent East-European Craton with values of 30-50·$10^{12}$ Pa·m, resulting from its high thermal age and associated low mantle temperatures and low surface heat flow values of 35-60 mW/m$^2$ [Kaikkonen et al., 2000; Moisio et al., 2000]. Other areas of high rigidity within Central and Western Europe are central parts of the North German Basin, the British Isles, the Rhenish, Armorican and Bohemian massifs, and the Iberian Massif. They represent thermally stabilized lithosphere of mostly Paleozoic thermal age with low heat flow values of 50-65 mW/m$^2$ (Figure 5.3b) and corresponding depth to the base of thermal lithosphere of 110-150 km (Figure 5.3c), which result in strength estimates in the order of 15-25·$10^{12}$ Pa·m (Figure 5.4).

A major axis of weakened lithosphere, caused by thermal perturbations with associated low thermo-tectonic ages, coincides with the Cenozoic Rift System (the Upper and Lower Rhine Grabens, Bresse Graben, extending into the Lower Rhone Valley extensional system [Ziegler and Dézes, 2005]. Similarly, the weakening of the lithosphere of Southern France can be attributed to the presence of tomographically imaged plumes rising under the Massif Central [Granet et al., 1995; Wilson and Patterson, 2001].

The weak lithosphere under the Pannonian Basin reflects another zone of profound Tertiary thinning of the thermal lithosphere. It is consequently characterized by extremely low strength [Lankreijer et al., 1999; Lenkey, 1999] that is in marked contrast with the high rigidity of the East-European and Moesian platforms. The low values for estimated integrated strength for the series of basins in the western portion of the North Sea including the Sole Pit are largely the result from the high surface heat flow input values of 70-
Integrated strength map for European lithosphere (reference model) overlain with the main tectonic features. Parameterization for model 1 is provided in Table 5.1. (a) Lithosphere strength lithosphere (b) Proportional upper crustal contribution to the total integrated strength. (c) Proportional mantle lithosphere contribution to the total integrated strength.
This study leaves open whether these heat flow values reflect overall higher lithospheric temperatures or perturbations due to blanketing effects and fluid flow in the sedimentary cover. Only in the former case, do high surface heat flow values correctly predict the strength. The locus in the North Sea of Mesozoic rifting and coeval thinning of the thermal lithosphere is represented by the Central and Viking Graben system. Owing to a long phase of post-rift lithospheric cooling [Pascal et al., 2002], surface heat flow has been reduced to estimates of 60-70 mW/m², implying a thermal lithosphere of 80 to 100 km thick. Moreover, due to stretching, crustal material has been replaced by stronger mantle material [van Wees and Stephenson, 1995; van Wees and Beekman, 2000; Cloetingh and van Wees, 2005]. Thus, these graben systems have regained part of their initial strength.

The Iberian Peninsula has also been affected by different thermo-tectonic phases with Variscan and Alpine compression and Mesozoic and Neogene (trans-) tensional tectonics. This polyphase history is reflected in pronounced variations in crustal thickness, surface heat flow and thickness of the thermal lithosphere. The lithosphere consequently exhibits large variations in lithospheric strength among the main tectonic domains. The Variscan inliers including the Iberian Chain and the inner massifs show high strength and are overprinted by Mesozoic and Cenozoic features, as for instance, the Ebro Basin and the Betics with lower rigidity. The perturbed heat flow and transitional crust of the Valencia Trough and Balearic Promontory caused by Mediterranean rifting and oceanization affect the calculated strength patterns as well.

The Adriatic microplate, with its high integrated strength forms a rigid indenter into the weakened NW-European Alpine foreland. The model predicts high strength of about \(20 \cdot 10^{12}\) Pa-m, as a consequence of thick and strong mantle lithosphere of low heat flow values of 45-60 mW/m², and moderate crustal thickness in this area. Local effects of meteoric fluid flow through highly fractured carbonate substratum further contribute to the anomalously low observed heat flow values. An accurate assessment of the rigidity of especially the Adriatic microplate is important for understanding the transmission of stresses into the neighbouring zones and the weakened Alpine foreland [Tesauro et al., 2005; Ziegler and Dèzes, 2007] and the Pannonian Basin system [Bada et al., 2006].

**5.4.4 Contribution from crustal and sub-crustal lithosphere**

In addition to the integrated strength map (Figure 5.4a), calculated over the entire lithospheric thickness, integrated strengths are also given for the upper and lower crust and mantle lithosphere, separately. Focusing more on variations than absolute values in strength, the proportional strength of the individual compositional layers is best given as ratios of the total integrated strength value. The resulting maps (Figure 5.4b and 5.4c) show that most of the bulk strength is in the mantle-lithosphere. The lateral strength variations are primarily caused by variations in the rigidity and temperature structure of the mantle-lithosphere. In contrast, variations in crustal strength appear to play only a secondary role. For instance, the strong lithosphere of the Fennoscandian Shield and Eastern European Craton, but also the Armorican, Bohemian and London-Brabant massifs, characterized by surface heat flow values less than 70 mW/m² and thermal ages in excess of 150 Ma, gain more than 80% of their rigidity from the cold and strong mantle (Figure 5.4c).

A marked increase in strength of the mantle lithosphere occurs across the Teisseyre-Tornquist Zone towards the Fennoscandian Shield. Within the overall strong Fennoscand-
The thick crust in the area of the Teisseyre-Tornquist Zone reduces the high integrated strength. The upper part of the cold and rigid mantle lithosphere is replaced by a particularly thick and weaker crust. The role of the compositional structure thus shows to be significant in second order strength patterns. Also in the Alpine domain, characterized by deep relative cool crustal roots, the crust contributes significantly to the thermal and rheological model setups outlining the definition of the key parameters for the thermal (Temp_mdl1) and rheological (rheol_mdl1) reference models and the parameters changed in each of the successive sensitivity models shown in bold.

### Table 5.2

<table>
<thead>
<tr>
<th>Sensitivity Tests</th>
<th>Reference Model (Figure 5.4)</th>
<th>Rheological Reference Model (Figure 5.4)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Composition</td>
<td>UC: quartzite, LC: diorite, ML: olivine</td>
<td></td>
</tr>
<tr>
<td>Temperature</td>
<td>Thermal model (Temp_mdl1 - reference) with heat production ($A_0$ in $\mu W/m^3$): sed: 1.0, UC: 2.0, LC: 1.0</td>
<td></td>
</tr>
<tr>
<td>Strain Rate</td>
<td>$10^{-15}$</td>
<td></td>
</tr>
<tr>
<td>Comments</td>
<td></td>
<td></td>
</tr>
<tr>
<td>'Cooler Model'</td>
<td>Same as in reference model</td>
<td></td>
</tr>
<tr>
<td>Temperature</td>
<td>Thermal model (Temp_mdl2) with heat production ($A_0$ in $\mu W/m^3$): sed: 1.0, UC: 2.0, LC: 1.5</td>
<td></td>
</tr>
<tr>
<td>Strain Rate</td>
<td>Same as in reference model</td>
<td></td>
</tr>
<tr>
<td>Comments</td>
<td>Higher heat production, lower mantle heat flow, cooler mantle lithosphere, higher integrated strength</td>
<td></td>
</tr>
<tr>
<td>'Warmer Model'</td>
<td>Same as in reference model</td>
<td></td>
</tr>
<tr>
<td>Temperature</td>
<td>Thermal model (Temp_mdl3) with heat production ($A_0$ in $\mu W/m^3$): sed: 1.0, UC: 2.0, LC: 0.5</td>
<td></td>
</tr>
<tr>
<td>Strain Rate</td>
<td>Same as in reference model</td>
<td></td>
</tr>
<tr>
<td>Comments</td>
<td>Lower heat production, higher mantle heat flow, warmer mantle lithosphere, lower integrated strength</td>
<td></td>
</tr>
<tr>
<td>'Diabase Model'</td>
<td>LC: diabase (instead of diorite)</td>
<td></td>
</tr>
<tr>
<td>Temperature</td>
<td>Same as in reference model</td>
<td></td>
</tr>
<tr>
<td>Strain Rate</td>
<td>Same as in reference model</td>
<td></td>
</tr>
<tr>
<td>Comments</td>
<td>Creep mechanisms for Diabase are activated at higher temperatures compared to Diorite thereby increase in strength of lower crust</td>
<td></td>
</tr>
<tr>
<td>'Low Strain Rate Model'</td>
<td>Same as in reference model</td>
<td></td>
</tr>
<tr>
<td>Temperature</td>
<td>Same as in reference model</td>
<td></td>
</tr>
<tr>
<td>Strain Rate</td>
<td>$10^{-17}$ s$^{-1}$</td>
<td></td>
</tr>
<tr>
<td>Comments</td>
<td>Viscous flow is easier activated for lower strain rates and integrated strength is lower from reduced viscous strength</td>
<td></td>
</tr>
</tbody>
</table>

Rheological and thermal model setups outlining the definition of the key parameters for the thermal (Temp_mdl1) and rheological (rheol_mdl1) reference models and the parameters changed in each of the successive sensitivity models shown in bold.
total integrated strength [Willingshofer and Cloetingh, 2003].

In contrast, the low rigidities of the Eifel and Massif Central volcanic provinces result from thermally weakened mantle lithosphere that correlates with tomographically imaged plumes [Granet et al., 1995; Wilson and Patterson, 2001]. The mantle lithosphere appear to be more severely thermally perturbed than the overlying crust so that mantle rigidity is almost completely absent while the crust retains some strength. Therefore, the strength of these thermally perturbed areas resides almost completely in the upper crust (Figure 5.4b). Over time, with the cooling of such lithosphere, the strength of the mantle lithosphere grows faster than the strength within the crust because temperature-dependant ductile flow is more important in the mantle. Also beneath the North Sea Basin a weakened mantle lithosphere (Figure 5.4c) is calculated, and can be attributed to destabilization processes occurring during the Permo-Carboniferous tectono-magmatic cycle [Ziegler et al., 2004] and again during Triassic to Early Cretaceous rifting [Ziegler, 1990; Ziegler and Cloetingh, 2004].

The calculated strength and ratios from the crustal and mantle contributions may vary as a result of model uncertainties. Especially the input temperature of the model contains uncertainties in crustal heat production and the amount of mantle derived heat. Increasing the average heat production in the crust, while keeping the surface heat flow gradient fixed, results in an overall cooler lithosphere because less heat is derived from the mantle (i.e. the ‘cooler model’; Table 5.2; Figure 5.5a). On the other hand, when the same surface heat flow is combined with a crust that produces less radiogenic heat, the heat flow from the mantle will be higher (i.e. the ‘warmer model’; Table 5.2; Figure 5.5a). Especially the colder mantle lithosphere of the ‘cooler model’ increases its rigidity and consequently gives higher integrated lithospheric strength. Inversely the ‘warmer model’ results in a reduction of the mantle and integrated lithospheric strength again especially for those areas that are the coldest in the reference models, whereas the change in already hot models is much smaller (Figure 5.5a).

The temperature and strength models in Figures 5.5b and 5.5c depict the effect of these lower and elevated temperature models on the calculated integrated strength. The areas, which in the reference model represent cold and strong lithosphere, are more strongly affected by alternative temperature models. Especially when combined with a thick crust, changes in the radiogenic heat production of the input model weight heavily on the predicted temperature structure and resulting rigidity of the upper mantle-lithosphere. This is for instance true for the Teisseyre-Tornquist suture zone, the Polish Trough and for the thickened crust further north under the Baltic Sea. Instead, areas that are already hot and weak in the reference model gain or lose almost any additional strength with the increase or decrease of the heat production.

Other variations to the specifications of reference model are a diabase instead of a diorite composition for the lower crust or the use of a lower uniform strain rate of for instance $10^{-17}$ s$^{-1}$ (Table 5.2). Such uniform changes on the model alter the relative crustal and mantle lithosphere contribution to the total integrated strength, but not the first-order lateral variations in strength.

5.4.5 Thickness and coupling of mechanically strong layers

A mechanically strong upper crust (MSUC), lower crust (MSLC) and mantle lithosphere (MSML) can be defined as layers of strength above 20 MPa with their respective thicknesses are labelled as $h_{uc}$, $h_c$ and $h_{ml}$ [Burov and Diament, 1995]. Mechanically strong
The effect of uncertainty in the temperature model on the calculated integrated strength. (a) Schematic illustration of the sensitivity of strength to temperature variations (b) The effect on strength from lower temperatures resulting from a higher crustal heat production. Increasing the crustal heat production $Q_{\text{crust}}$ while keeping $q_{\text{surf}}$ (surface heat flow) fixed results in a reduced $q_{\text{mantle}}$ and therefore cooler and stronger (mantle) lithosphere. (c) The effect on strength from elevated temperatures resulting from a reduced crustal heat production. Lower crustal heat production with same surface heat flow, increased heat flow contribution from mantle, what leads to a warmer and weaker (mantle) lithosphere.
layer thicknesses can be derived from the 3-D strength model for the different compositional layers (Figure 5.6). The mechanical strength of the upper crust ranges between 4 km for the thermally weakened regions (e.g. Eifel and Massif Central volcanic provinces and Pannonian Basin) up to 16 km for the strong stable Craton. With a lower crust characterized by a diorite rheology, ductile creep laws are relatively easily activated so that the lower crust is weak. Mechanical strength in the dioritic lower crust is only present where temperatures are very low (Figure 5.6b). This condition can be either fulfilled in areas of thinned crust with recalibrated crustal temperatures as found in old rifts and passive margins or in stable cratonic regions where thick crust is combined with extremely low crustal temperatures. A mechanically strong lower crust is found along the European western passive margin and beneath the cooled Central and Viking grabens and Rockall Trough. The MSLC for the East European Craton, Fennoscandian Shield, and Armorican and Bohemian massifs results from extremely low crustal temperatures despite their thicker crust. Strong lower crust also occurs in the regions with Paleozoic thermo-tectonic ages and moderately crustal thickness less than 35 km.

The mantle lithosphere exhibits significant mechanical strength. For the Paleozoic inliers (e.g. London-Brabant and Bohemian massifs) show thickness estimates of 25-35 km while the Fennoscandian Shield and EEC show values of 60 km and higher. The mechanically strong mantle lithosphere is also extremely thin or completely absent over large areas where mantle temperatures are high (e.g. the Eifel and Massif Central volcanic provinces). Furthermore areas with high crustal thickness (like the orogenic crusts of the Alps) lack a mechanically strong mantle lithosphere, although the great uncertainty of the mantle temperatures calls for great caution here. Average mechanically strong upper mantle thicknesses of 15-30 km occur over large areas of Europe.

The occurrence of coupling between the mechanically strong layers is an important mechanical feature and depends on variations in composition, thickness of the crust and the temperature distribution [Stephenson and Cloetingh, 1991; Burov et al., 1993]. Coupling between the mechanically strong levels of the different compositional layers occur when the strength in the weak intervals remain above the threshold of 20 MPa. The coupling between the mechanically strong layers is controlled by the extent to which ductile creep is activated in the lower part of each compositional layer. Especially the strength of the lower crust is critical for coupling layers of mechanical strength between the crust and mantle. Ductile creep dominates the lower crust for a diorite rheology and average temperature conditions.

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5.4.6 Conclusions and discussion

The final step in the assessment of the European lithosphere strength comes from estimates of the effective elastic thickness (EET). Following Equations 5.6 and 5.7, EET predictions can be made from the derived distribution in thickness and coupling of the involved mechanically strong layers. As expected, areas of high EET correspond to regions of high integrated strength. The high EET values for the East European Craton and Europe’s Paleozoic inliers reflect the high mechanical strength of the mantle part of the lithosphere. Areas with low EET such as the Pannonian Basin reflect the weak mantle.

These results can be compared with EET estimates from Bouguer coherence and free-air admittance analyses [Pérez-Gussinyé and Watts, 2005]. Both gravity data analyses and the forward rheological approximation of their study show a high flexural rigidity in the Eastern European Craton. Differences between the studies occur for the European lithosphere westward of the Teisseyre-Tornquist Zone, where the long wavelength spatial patterns inferred from their analysis show less resemblance to first order tectonic features of Europe. For example, the mechanically strong Eastern European Platform extends across
the Teisseyre-Tornquist Zone, showing high values for the North German Basin and the southern North Sea. Both methodologies have limitations as discussed above for our approach and in Pérez-Gussinyé and Watts [2005] for the use of gravity data for lithosphere rigidity. However, both estimates point clearly to pronounced spatial variation in mechanical properties.

The modelling significantly generalizes the compositional and thermal input, accounting for first order structures and neglecting local, second order scale deviations. As result, the rheological structure of the European lithosphere deviates from our first order 3-D model, and consequently will affect the estimated integrated strength.

For instance, local variations in crustal composition and architecture (e.g. caused by faults displacing parts of the crust) were not incorporated in our model. Understandably, strongly deformed areas like the Alpine and Pyrenees orogenic zones comprise a substantially thickened and complex crustal architecture that is difficult to represent in a first-order compositional layer model. An important factor is also presence of water or serpentinitization of the subcrustal lithosphere, both of which reduce the strength of the mantle [e.g. Pérez-Gussinyé and Reston, 2001], or melt depletion, which strengthens the mantle [van Wijk and Cloetingh, 2002]. On the contrary, zones of eclogitized crustal material in orogens, can potentially weaken the strong upper part of the mantle–lithosphere [see also Willingshofer and Cloetingh, 2003]. These factors, not accounted for in the models, contribute to weakening of former mobile zones, preferentially reactivated during the breakup of Pangea [e.g., Janssen et al., 1995].

The limitation of the temperature input model is that it assumes first-order steady-thermal conditions and the uncertainty to represent properly the boundary condition at the lithosphere base. Steady temperature conditions are disturbed by active tectonics, in the form of burial and erosion, when this results in fast effective vertical motion. Lastly, the model assumes a uniform bulk strain-rate (i.e. $10^{-15}$ s$^{-1}$) that is considered representative for an intraplate setting, but may deviate in active neotectonic zones as well as areas close to plate boundaries.

The strength model can be further improved by better constraints on the composition and structure of the lithosphere, e.g. from the refined EuCrust-07 compositional model [Tesauro et al., 2008], or by generating a temperature input model by solving the heat equation in fully 3-D and with an thermal boundary condition at the effective irregular lithosphere base. However, certain limitations to classical strengths are intrinsic. One restriction is that only bulk isotropic strength are quantified and mechanical anisotropies or discrete strength reduction are neglected (i.e. preferred crystal orientation, foliation, fractures and fault zones [e.g. Dirkzwager et al., 2000]). Strength is direction-dependent as a result of folding and foliation of rocks, affecting the localization of deformation along pre-existing tectonic fabric. Rheological studies of Phanerozoic basin evolution demonstrate that the interplay of stresses and tectonic fabric make them permanently prone to reactivation [Dirkwzager et al., 2000; van Wees and Stephenson, 1995].

Meanwhile, the strength-map compilation of the European lithosphere (case I in this study) describes well the first-order rheological discontinuities that arise at compositional and thermal lithosphere transition zones. Major contrasts in mechanical strength exist between the thermally stable lithosphere of the East-European Platform, east of the Teisseyre-Tornquist Zone, and the relatively weak lithosphere of major portions of thermally perturbed central-western Europe. The high strengths inferred for the East-European Platform and Fennoscandian Shield reflect low mantle temperatures and sur-
face heat flow. Regions of high rigidity further occur at the site of thermally stabilized Paleozoic massifs (e.g. Rhenish, Armorican, Bohemian, and Iberian massifs) characterized by low heat flow and high lithospheric thickness values. In contrast, a major axis of low lithospheric strength is found along the European Cenozoic Rift System (e.g. Rhine Graben, and Massif Central) and can be attributed to the presence of tomographically imaged plumes. Although substantial plume-related thermal thinning of the mantle lithosphere is confined to the Massif Central and the Eifel area, the mantle lithosphere has been thermally weakened in a broad zone around the European Cenozoic Rift System.

Another pronounced contrast in strength is found between the weak lithosphere of the thinned and hot Pannonian back-arc system and the strong surrounding lithosphere of the Adriatic Plate and Moesian Platform. These lateral variations in lithosphere strength likely played a significant role in the lithosphere flexural response [Leever et al., 2006]. Spatial transitions in lithosphere rigidity play a noticeable role on the temporal evolution of a foredeep during orogenic loading localizing and amplifying the flexural response to the rheological transition. The inferred strength distribution provides a framework for better understanding of the first-order deformation patterns in Europe’s intraplate lithosphere.

5.5 Case 2: lithospheric structure of SW Canada

The Canadian Cordilleran deep structure has been intensively studied over the past decades with high quality geophysical probing. After having successfully built the workflow with the European strength map, the prolific geophysical dataset for the south-western Canadian lithosphere allows for a comparable attempt. The work by Lowe and Ranalli [1992] on the temperature and rheological structure provides a valuable precedent for this current study. The main focus here is the transition in the lithosphere from the south-eastern Canadian Cordillera to the stable North American Craton in the east, between which several crustal and upper mantle geophysical parameters change [Hyndman et al., 1999].

5.5.1 Compositional structure

Seismic refraction and reflection studies [Kanasewich et al., 1994; Burianyk and Kanasewich, 1995; Cook, 1995; Clowes et al., 1995] have elucidated the crustal and lithospheric structure of the study area beneath the south-eastern Canadian Cordillera. The map of the depth to the Moho (Figure 5.7a) exhibits an area beneath the FFTB where the crust reaches high thickness of 46-50 km [Burianyk et al., 1997]. The most prominent structure on the Moho in the southern Canadian Cordillera is observed near the Rocky Mountain Trench, where the crust thins westward of the FFTB by about 10 km [Cook, 1995]. In the Omineca Belt the crust exhibits a thickness of 32-36 km and is considered the result of strong post-orogenic crustal thinning due to core complex formation [Parrish et al., 1988; Cook et al., 1992]. East of the FFTB, the crustal thickness reaches average cratonic values of 38-42 km (Figure 5.7a) with variations related to the regional presence of a thick sedimentary sequence (e.g. crustal thickness exceeds 45 km beneath the Western Canadian Sedimentary Basin depocenter).

The Moho discontinuity can be imaged through geophysical probing mainly due to the density change that occurs from a quartz-dominated to an olivine-dominated composition. Several decades of extensive reflection and wide-angle-refraction seismics studies
combined with alternative geophysical probing (i.e. aeromagnetic and gravity data) thus provide detailed imaging of the crustal and also lithosphere structures [Clowes et al., 1995; Zelt et al., 1996; Hope and Eaton, 2002; Ross, 2002]. For thermal and rheological modelling, not only the density but also alternative rock parameters are required (like thermal conductivities, distribution of radiogenic elements and rock creep parameters). Although these parameters, like for rock density, are dependant on rock composition, their variability does not necessarily correlate positively to density variations.

Seismic refraction lines across the Western Canadian Sedimentary Basin reveal distinct basement blocks [Clowes et al., 2002] of different Archean-Proterozoic tectonic origin. The Alberta basement comprises Proterozoic fragments of granulite composition, and to the southwest the Archean Loverna and Medicine Hat blocks are divided by the Vulcan structure [Clowes et al., 2002]. As alternative to these geophysical interpretations from density variations, xenoliths provide direct rock samples from depth brought to surface at igneous provinces. The Sweet Grass Hills forms an Eocene igneous complex in the central Alberta basement that contains (para-)gneisses and mafic granulites. They are respectively derived from middle to lower crustal levels within the Medicine Hat block [Ross et al., 1991]. Also the Loverna block presumably consists of paragneisses. The intervening Vulcan structure is considered as being related to an Archean subduction zone interpreted from a few amphibolite basement samples. The crust of the Medicine Hat block thickens southward to more than 55 km thickness, which may be explained by magmatic underplating as inferred from seismic velocity refraction data together with mafic granulite xenoliths [Ross et al., 1991; Ross, 2002].

Control on lithospheric structure in south-western Canada (Alberta-British Colombia) from (a) crustal thickness distribution [Burianyk et al., 1997] and (b) near-surface heat flow constraints with basement heat flow from BHT in basin [Bachu, 1993] and heat flow constraints from the Cordillera [Majorowicz and Gough, 1991; Lewis et al., 1992]. For orientation, note the US-Canada Border (base), and borders of British Colombia-Alberta (centre) and Alberta-Saskatchewan (right) provinces. The projected coordinates are in meters according to the GCS_North_American_1983 with NAD_1983_Albers projection.
Compilations for the south-western Canadian lithosphere. (a) Interpolated surface heat flow distribution as obtained from constraints in Figure 5.7b. (b) Sediment thickness distribution [Bachu, 1993]. (c) Tectonic domains [Clowes et al., 2002] with (1) the Laverna Block, (2) Vulcan Structure and (3) Medicine Hat Block. (d) Interpolated crustal thickness distribution as obtained from constraints in Figure 5.7a. (e) Calculated mantle heat flow distribution. (f) Calculated thermal lithosphere thickness map.

5.5.2 Constraints on crustal geothermal field

Direct observations on the geothermal field can only be acquired by measuring temperature gradients and rock thermal conductivities from drill holes that normally only reach shallow crustal depths. Especially the foreland basin domain provides a dense coverage of bottom-hole temperatures (BHTs) from petroleum wells that provide an extensive heat flow data repository [e.g. Majorowicz and Jessop, 1981, 1993]. The prolific petroleum well drilling provides for abundant BHT measurements. However, the scarcity of in-situ thermal conductivity measurements and limited depth ranges hamper successful thermal field estimation [Hyndman and Lewis, 1999; Majorowicz et al., 1984; Majorowicz and Jessop, 1993]. Figure 5.7b depicts heat flow contour maps derived from the dense coverage of BHTs as examined by Majorowicz et al. [1984], Majorowicz and Jessop [1993] and Bachu [1993]. These compilations of temperature gradients and heat flow estimates
at the base of the sedimentary cover give overall low geothermal gradients of approximately 25-30 °C/km. With estimated conductivities for the sediments in range of 1.6-2.3 W·m⁻¹·K⁻¹, the thermal gradients correspond with low heat flow values of 40-55 mW/m² [Bachu, 1993; Majorowicz et al., 1984; Majorowicz and Jessop, 1993] for the basin (Figure 5.7b).

A few detailed studies on a series of wells from the Cordillera interior (Intermontane and Omineca belts) provide reliable geothermal field constraints as they contain in-situ thermal conductivities and radiogenic heat-production measurements against depth [Lewis et al., 1992]. Borehole measurements in the Purcell Range for instance, located just west of the Rocky Mountain Trench that borders the FFTB, give heat-flow estimates of 92-99 mWm² (Figure 5.7b) and conductivities of 3.5-4.5 W·m⁻¹·K⁻¹. Average heat generation values for the Purcell Range are about 3.6 μW/m³, but values as high as 5-6 μW/m³ are found for Early Cretaceous intrusives [Lewis et al., 1992 and references therein]. Heat flow values in the Intermontane and Omineca belts are notably high. For the Omineca Belt, surface heat flow ranges between 70-90 mW/m², but also contains values as high as 120 mW/m² [Majorowicz and Gough, 1991; Lewis et al., 1992; Hyndman and Lewis, 1999]. Both high crustal heat generation from plutonic rocks of considerable vertical extent and elevated mantle heat flow contribute to these heat flow estimates for the Cordillera interior. Strong variations in surface heat flow can be correlated to differences in crustal heat generation from the different plutonic rocks (10-60 mW/m³), while the regionally high heat flow is explained with an overall strong mantle flow of ~64 mW/m² [Lewis et al., 1992].

The separation of crustal and mantle heat contribution is more difficult to assess for the FFTB and the basin. A few samples from the basement beneath the basin give averages around 1-3 μW/m³, but with a large variability from less than 1 μW/m³ to more than

<table>
<thead>
<tr>
<th>Material</th>
<th>C&amp;T87</th>
<th>R&amp;M87</th>
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<tr>
<td>Quartzite</td>
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<tr>
<td>Granite</td>
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<td>Dunite</td>
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<tr>
<td>Olivine</td>
<td></td>
<td></td>
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<tr>
<td>Mafic-granulite</td>
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5 Strength patterns of intraplate lithosphere

8 μW/m³ [Majorowicz and Jessop, 1981].

The contribution of heat generation from the basin sediments is generally considered significantly less than for the crystalline basement, but still gives values of 0.4-1.6 μW/m³ [Lewis et al., 1992]. Varying heat production from the sediments with a thickness of 5 km can still contribute to ~10 mW/m² in observed heat flow depending, on clay and feldspar content [e.g. Bachu and Burwash, 1991]. Crystalline basement rocks of the Craton are only exposed further north-eastward in the central Canadian Shield. Here heat-flow estimates vary around 40-50 mW/m² with spatial variability correlated to elevated radioactive heat production from plutons [e.g. Guillou-Frottier et al., 1996]. Whether comparable variations in heat flow from the Cordilleran foreland basin may also be ascribed to variations in heat production of the underlying crust or whether the basin sediments themselves induce the observed perturbations in the (near surface) heat flow distribution has been debated [Bachu, 1993; Majorowicz and Jessop, 1981; Majorowicz et al., 1984]. Meanwhile, there is an overall agreement on a more uniform thermal mantle signature, what permits to treat it as a single thermo-tectonic province in the present study.

5.5.3 Large-scale lithosphere thermal structure

Modelling of the thermal structure of the lithosphere is described as a simple steady-state heat-diffusion problem for which interpolated surface heat-flow and temperature gradient contour maps are used as upper boundary condition (Figure 5.8a). A sedimentary thickness map describes the infill of the foredeep basin (Figure 5.8b) and two crustal layers are defined from the crustal thickness map (base to upper crust is half of the depth to the lower crust) (Figure 5.8d). The compositional layers involve conductive heat parameters and radiogenic heat production. The compositional and temperature modelling combines these constraints (Figure 5.7) for calculating vertical temperature profiles from the conductive heat equation (Equation 5.1) and estimating the total amount of crustal heat production and mantle derived heat flow (Figure 5.8e).

The Intermontane and Omineca belts (Figure 1.4) are generally considered as a single heat flow province with elevated surface heat flow resulting from high mantle heat flow as result of a strong attenuation of the thermal lithosphere [Gough, 1986; Lewis et al., 1992; Hyndman and Lewis, 1999]. The regional temperature structure for the Omineca Belt yields average surface heat-flow values of ~80-90 mW/m² that corresponds with a thermal gradient in the order of 30-35 °C/km for the upper crust. The modelling estimates a crustal heat production of 20-40 mW/m² and predicts a mantle heat flow in the order of 55-65 mW/m² (Figure 5.8e) for the thinned, presumably granitic crust. Given the high surface heat flow and calculated heat production, the conductive heat equation gives temperatures up to 600 °C at 20 km depth and a 1300 °C isotherm (i.e. base thermal lithosphere) at 60-90 km depth (Figure 5.8f).

These trends can be compared to inferences from geophysical imaging of electrical conductivities, magnetotelluric and seismic velocities that are considered to be an indirect proxy of the lower crustal and mantle temperature structure [Gough, 1986; Jones and Gough, 1995; Hyndman and Lewis, 1995; Clowes et al., 1995]. The Omineca Belt is characterized by high electrical conductivities in the middle to lower crust and a pronounced mantle low-velocity zone at a depth of only 50+/−5 km. Generally, high heat flow and resulting high crustal temperatures may account for overall low seismic velocities as is observed along a NS trending axis of attenuated lithosphere throughout the Cordil-
lera [Hyndman and Lewis, 1995; Clowes et al., 2002]. Hyndman and Lewis [1995, 1999] provide an insightful cross-Cordilleran compilation along the geophysical Lithoprobe transect, integrating direct upper crustal temperature measurements and geophysical remote-sensing studies on the deeper lithospheric thermal structure. Their compilation clearly highlights the first order trend of substantially thinned lithosphere beneath the Intermontane and Omineca belts, and typical cratonic lithosphere beneath the North American Craton, thus pointing to a significant thermal transition beneath the FFTB.

The temperature structure of the Craton to the east stands in strong contrast. The model exhibits mantle heat flow as low as 15 mW/m² derived from a surface heat flow of 40-55 mW/m² and 30-40 mW/m² of crustal heat production (Figure 5.8e). It has already been mentioned how the temperature structure of the basin, derived from bottom-hole well measurements, and of the underlying basement have been strongly debated [Bachu, 1993; Majorowicz and Jessop, 1981; Majorowicz et al., 1984]. Uncertainty exists around the conversions from bottom-hole temperatures to heat flow, due to the lack of in-situ thermal conductivity measurements. For the temperature modelling, firstly converted basement heat-flow values were adopted from Bachu [1993]. The calculated thicknesses of the thermal lithosphere of the craton reached values well above 200 km. Alternatively, the original temperature gradient maps for the sedimentary cover can be used as boundary condition. With temperature gradient variations of 25-35 °C/km and adopting a conductivity of 1.8-2.0 W·m⁻¹·K⁻¹ result in lower thermal lithosphere thickness estimates. Thermal thicknesses of 120-200 km for the Craton (Figure 5.8f) comply better with earlier estimates [Lowe and Ranalli, 1993; Hyndman and Lewis, 1999].

The modelling signifies a thermal contrast between the lithosphere thickness of the Omineca Belt and Craton of less than 80 km for the first to well above 160 km for the latter. The modelling has difficulty outlining the transition due to differences in thermal input between the two thermal provinces and insufficient data coverage for the FFTB in between. The transition presumable covers a ~200 km wide zone [see also modelling results from Lowe and Ranalli, 1993]. Each of the two thermal provinces also exhibits lateral variability in the thermal modelling results (Figure 5.8f). For the Cordillera, better mapping of large plutons would be required to discern in how far heat flow values above 80 mW/m² may be attributed to locally high radiogenic heat production. For the basin, lateral variability in conductivity and sedimentary thicknesses would need to be better mapped. Lateral variations in crustal heat production are for the current model setup mainly due to variations in crustal thickness. A thick crust increases the crustal contribution to the measured surface heat flow and consequently reduces the mantle heat flow contribution. Moderate heat flow values with a thick crust imply low mantle heat flow and consequently a cold mantle lithosphere, as shown under the central part of the FFTB (see Figure 5.8d and Figure 5.8f for crustal and thermal lithosphere thickness).

5.5.4 Rheological variations from compositional variability

The viscous flow parameters that are determined in rock deformation experiments outline how the resistance to viscous flow is low from the presence of quartz minerals [e.g. Carter and Tsenn, 1987]. The resistance viscous flow increases with the loss of water, a reduction in felsic components, and an increase in Mg/Fe ratio (i.e. by partial melting and depletion of residue).

For the crust, granitic rocks are weakest as they form aggregates of feldspar and quartz and have high water content. On the other hand, dry quartzite is the upper-crustal
rock type that is most resistant to viscous flow as it lacks the contribution of feldspar and water. The change of viscous strength can be plotted for different rock types as function of temperature and assuming a given temperature profile and strain rate of $10^{-15}$ s$^{-1}$ (Figure 5.9). The numerous rock deformation experiments have illustrated the variations in creep parameters for different rock samples of similar rock types. For instance, the dry quartzite of Ranalli and Murphy [1987] is among the strongest, whereas the dry quartzite from Carter and Tsenn [1987] falls in between the former and wet granite (Figure 5.9). The dry quartzite of Ranalli and Murphy [1987] will be used for the upper crust of the Canadian Craton (case II), compared to the weaker dry quartzite from Carter and Tsenn [1987] that was used for the upper crust in the European strength-map (case I).

The contribution of more mafic constituents in the crust (especially the lower crust) will further increase the resistance to flow under the same temperature conditions. The diorite of Carter and Tsenn [1987] and quartz-diorite of Ranalli and Murphy [1987] are notably stronger than the more felsic rock types (Figure 5.9). However, the increase in flow resistance for lower crustal rocks is typically insufficient to compensate for the much higher lower crustal temperature (~400-600 °C) what makes the lower crust still prone to viscous flow. A diabase composition [Carter and Tsenn, 1987; Ranalli and Murphy, 1987] adds resistance and only mafic granulites [e.g. Wilks and Carter, 1990] will hamper lower crustal flow significantly.

It has been previously stated that lithospheric bulk strength is especially controlled by its temperature structure and only second by the compositional heterogeneity. This is...
true particularly because of the mantle lithosphere that normally makes more than 70% of the total integrated strength (see for instance the European lithosphere; Figure 5.4c) and for which strength differences originate especially from temperature variations. Instead, for the crust itself, a change from diorite to diabase composition of the lower crust results in an increase of the viscous strength that is similar to a 100 °C temperature reduction. For the European lithosphere (case I), a change of the lower crust from a diorite to diabase composition resulted in a notable increase in lithosphere strength for those areas that lack mantle strength due to high temperatures. However, as outlined for the Canadian cratonic basement, the compositional structure may comprise blocks of different composition and future strength-map compilations may account for this.

For the upper mantle-lithosphere, many studies have addressed the viscous flow parameters for a variety of olivine aggregates (e.g. olivine, dunite or basalts) [Goetze and Evans, 1979; Chopra and Paterson; 1984; Hirth and Kohlstedt, 1996]. The absence of felsic components give these aggregates overall high resistance to flow, and variations in flow parameters mainly occur due to depletion of Fe relative to Mg components and especially because of the amount of water [Chopra and Paterson; 1984; Hirth and Kohlstedt, 1996]. For the European strength-map, the creep parameters for olivine from Goetze and Evans [1979] were used. The viscous strength as function of temperature is similar for the dry olivine from Hirth and Kohlstedt [1996] and dry dunite from Carter and Tsenn [1987] (Figure 5.9). The first has a low water content and dunites are depleted in Fe components.

<table>
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<tr>
<th>Mineral/Rock</th>
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<th>Dorn Law</th>
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<tr>
<td></td>
<td>n</td>
<td>$E_p$ (kJ-mole$^{-1}$)</td>
</tr>
<tr>
<td>wet granite [Ranalli and Murphy, 1987]</td>
<td>1.9</td>
<td>137</td>
</tr>
<tr>
<td>wet quartzite [Ranalli and Murphy, 1987]</td>
<td>2.3</td>
<td>154</td>
</tr>
<tr>
<td>wet quartzite [Gleason and Tullis, 1995]</td>
<td>4.0</td>
<td>223</td>
</tr>
<tr>
<td>dry quartzite [Ranalli and Murphy, 1987]</td>
<td>2.4</td>
<td>156</td>
</tr>
<tr>
<td>quartz diorite [Ranalli and Murphy, 1987]</td>
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<td>219</td>
</tr>
<tr>
<td>diabase [Ranalli and Murphy, 1987]</td>
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<td>260</td>
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<tr>
<td>Olive [Goetze and Evans, 1979]</td>
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<tr>
<td>wet olive [Hirth and Kohlstedt, 1996]</td>
<td>3.5</td>
<td>498</td>
</tr>
<tr>
<td>dry olive [Hirth and Kohlstedt, 1996]</td>
<td>3.5</td>
<td>515</td>
</tr>
</tbody>
</table>

Rheological model parameters for Canadian Cordilleran and Craton crust and mantle (case II).
For the strength-map compilation of the south-western Canadian mantle lithosphere the dry olivine from *Hirth and Kohlstedt* [1996] is chosen, although the olivine according to *Goetze and Evans* [1979] or the dry dunite from *Carter and Tsenn* [1987] will give similar results. As discussed in Chapter 6, the geodynamic modelling tests the implications of a compositional change from dry olivine for the cratonic mantle lithosphere against a wet olivine *Hirth and Kohlstedt* [1996] for the mantle lithosphere of the Cordillera interior. The latter is considered containing significant amounts of water, derived from melting of a subduction slab beneath the backarc.
5.5.5 Calculated strength-map

The rheological model of the south-western Canadian lithosphere (i.e. the lithosphere of SE Cordillera and the easterly North American Craton) is built applying a work flow similar to the one used in case I (Figure 5.2). A uniform layer cake model is applied as compositional input, comprising an upper and lower crust of dry quartzite [Ranalli and Murphy, 1987] and diabase [Carter and Tsenn, 1987], respectively, and a mantle lithosphere with dry olivine composition [Hirth and Kohlstedt, 1996] (Table 5.3). Brittle and ductile rheological laws are adopted (equations 5.6, 5.7 and 5.8and rheological parameters assigned to the compositional layers (Table 5.1). The 3-D strength distribution is calculated and integrated over the depth range, and the resulting integrated strength-maps for crust and lithosphere are given in Figures 5.10a and 5.10b.

Both the integrated crustal and lithosphere strength-maps elucidate the pronounced rheological contrast between the weak Cordillera and the mechanically strong Craton lithosphere. This rheological transition in the lithosphere resides beneath the FFTB, which forms an upper-crustal transition between the strongly deformed Cordillera interior and the non-deformed foreland basin sediments. The rheological transition is the result of the strong thermally perturbed mantle beneath the Cordillera [Hyndman et al., 2005]. The hot and consequently weak Omineca Belt exhibits integrated crustal strengths of ~1.0·10³ GPa·m and only 0.1-1.0·10³ GPa·m for the mantle lithosphere. The integrated strengths for the Craton are more than one order of magnitude higher with 4.0-10.0·10³ and 10.0-40.0·10³ GPa·m.

A lithospheric section across the rheological transition from the Cordillera to Craton reveals the lateral variation in depth-dependant strength (Figure 5.10c). Brittle strength increases with depth to the point where the strength is limited by ductile flow which strength decreases with depth. The changes between the compositional layers alter the ductile strength as the mineral composition sets different creep parameters. For the Cordillera, strength mainly resides in the upper crust as the lower crust and mantle lithosphere are severely weakened by high temperatures [Figure 5.10d]. The lower crust and mantle lithosphere only reach strengths of 40-60 MPa and the mechanically strong crust is detached from the mantle lithosphere. For the Craton instead, the mechanically strong crust and mantle lithosphere layers reach strengths of 0.6-2.0·10³ MPa and are welded together (Figure 5.10c), accounting for 70% of the total integrated strength (see for instance the European lithosphere; Figure 5.4c). Such a pronounced rheological transition likely affects the geodynamic response when imposed to large tectonic forces. The thermal attenuation presumably only came in existence in Eocene time after shortening in the Cordillera and FFTB to the east ceased [Lowe and Ranalli, 1993].

Lateral variations compositions of the crustal and mantle layers may occur. Large scale magmatism likely affected the Omineca Belt at regional scale and also distinct crustal blocks have been deciphered for the cratonic basement. The mantle lithosphere of the Omineca Belt may be further weakened by the presence of water from subduction related melts [Currie et al., 2004, 2008]. The Craton instead likely comprises a dry and depleted mantle composition.

5.5.6 Rheological heterogeneities

Estimates of thicknesses of the mechanically strong (> 20 MPa) upper crust (MSUC), lower crust (MSLC) and mantle lithosphere (MSML) [Burov and Diament, 1995] are derived from the 3-D strength model. The above discussed rheological cross-sections show
Strength patterns of intraplate lithosphere particularly well the stratification in mechanical strength. The base of the MSML is depicted (dashed line) and yields a thickness of only 10-20 km at the strongly attenuated lithosphere of the Omineca Belt whereas the mantle lithosphere of the Craton exhibits a mechanically strong layer thickness of around 175 km.

Mechanical strength in the upper crust is limited by the depth-dependant brittle failure but easily reaches strength values beyond the mechanical strength threshold. The mechanical strength of the lower crust is more strongly affected by the activation of ductile creep which is mineral and temperature dependant. For a diorite lower crust, temperatures above 600 °C activate ductile flow by which it loses mechanical strength. Decoupling between the mechanically strong crust and mantle may consequently occur. For the Cordilleran lithosphere, the crust may comprise a mechanically strong layer of 10-20 km, underlain by a hot and weak mantle lithosphere with a mechanically strong layer thickness of another 10-20 km (Figure 5.10c). These layers are decoupled by a hot and weak lowermost crust. This is different for the Craton (Figure 5.10c), where also the lower crust is entirely mechanically strong (> 30 km) and welded to a thick mechanically strong mantle lithosphere.

The effective elastic thickness distribution is the sum of these mechanically strong layers including the effect of coupled and decoupled rheology [Burov and Diament, 1995]. High EET values for the Craton are derived from the thick mechanically strong mantle lithosphere coupled to the mechanically strength in the crust. EET thicknesses amount to 80-120 km (Figure 5.10e). For the Cordillera instead, EET only reaches thicknesses of 20-40 km (Figure 5.10e) due to an almost absent mechanically strong mantle layer.

Traditionally, EET estimates were derived from the flexural foreland basin response to an orogenic load. The studies on the Canadian Cordilleran foreland basin included refinement from simple elastic plate approximation by accounting for a coupled viscoelastic response [Beaumont, 1981] and considering lateral variations in EET [Wu, 1991]. Analyses on the deflection pattern of the base of the foreland basin infill give best results for models that contain a lateral EET transition with 38 km west of the FFTB to 200 km beneath the North American Craton [Wu, 1991].

Alternatively, EET estimates may be obtained from coherence analysis of Bouguer gravity and topography Flück et al. [2003]. Comparison with these EET estimates reveals a similar trend. Flück et al. [2003] estimates a 10-30 km thick effective elastic plate for the Cordillera thickening to more than 80 km for the Craton lithosphere. The largest EET gradient is found approximately at the deformation front. EET estimates from this strength-map compilation give overall higher EET values when compared to estimates from gravity-topography coherency studies. Translation of lateral variations in near surface temperature constraints to mantle depths tends to overemphasize discontinuities. On the other hand, EETs estimated from gravity-topography coherency studies underestimate EET, because the longest wavelengths in the spectrum are not well resolved [Flück et al., 2003].

The EET estimates in this study are derived as a function of surface heat flow, crustal thickness and heat production, and rheologies of the respective layers. The spatial variations in each of these parameters determine the predicted EET distribution (Figure 5.10e). The relevance of these factors on EET variations can also be illustrated in a parameter-space analysis. Surface heat-flow input is iterated over 30-90 mW/m², heat production over 1.0-5.0·10⁶ W/m³, and crustal thickness over a 20-60 km interval. The resulting distribution in EET is shown for slices through the parameter-space with surface
heat flow against crustal thickness (Figure 5.11a) for a crustal heat production of $1.5 \cdot 10^{-6}$ W/m$^2$ and crustal heat production against surface heat flow (Figure 5.11b) for a crustal thickness of 35 km.

Like for the strength maps, the EET exhibits a first-order contrast in the rigidity of the lithosphere with values between 10 to 160 km. Low values originate from weak and hot lithosphere from high surface heat flow in combination with high heat production in thick crust. High EET values of more than 80 km are given for lithosphere with a relative surface heat flow and crustal heat production. The crustal thickness itself has a secondary effect (Figure 5.11a) unless the crustal heat production is high (Figure 5.11b).

EET estimates also provide for approximation of flexural wavelength assuming a simple 2-D elastic plate under a point load (Equation 5.9). High EET values of 80 km or higher result in flexural wavelengths above 500 km (Figure 5.11). This applies to the cold and strong lithosphere of the Canadian Craton. The weak lithosphere of the Cordillera instead, with EET values less than 40 km, yield flexural wavelengths less than 300 km. Intermediate lithosphere (see earlier case I western Europe) with surface heat flow of 60-80 mW/m$^2$ and thermal thickness of 100-150 km give EET values of 40-80 km and corresponding flexural wavelengths of 350-500 km.

5.6 Conclusions and discussion

The ability of the strength-map compilations to image the first order strength patterns of large lithosphere domains is well illustrated with the description of first-order rheological heterogeneities in the NW European (case I) and SW Canadian (case II) lithosphere. The 3-D strength model provides for maps that depict integrated strength and thicknesses of mechanically strong layering along with EET estimates and rheological cross-sections.

For case I, these results depict the heterogeneous distribution of strength for the European lithosphere, reflected in its polyphase deformation. The model predicts a patchwork of high strength and low strength domains in the Europe’s intraplate lithosphere that results from variations in composition and thermal regime, and can be tied to structural and thermal discontinuities. The European strength map reveals first-order rheological discontinuities between cratonic and thermally perturbed lithosphere.

Also for the south-western Canadian lithosphere, a substantial rheological discontinuity has been described by this study. Again a profound thermal contrast that occurs between the hot Cordillera and cold Craton determines the contrast in integrated strength and thickness of the mechanically strong mantle lithosphere.

Depending on the nature of the thermal and rheological discontinuity, the geodynamic implications may affect the style, focus and timing of large strain patterns at the rheological transition between Cordillera and Craton. The discontinuity was likely emplaced after post-orogenic rigorous thinning of the crust and lithosphere [Ranalli et al., 1989; Carr, 1992; Bardoux and Mareschal, 1994]. The here described (Tertiary) rheological discontinuity has not been the cause but the outcome of thinning of orogenic crust and lithosphere. The thinned crust and mantle lithosphere at the Omineca Belt forms a post-orogenic feature. The thicker crust in between the Omineca and FFTB likely forms a remnant of the orogenic crust. Here, the lithosphere encompasses unstretched crust and surface heat flow is presumably low due to its location on the edge of the undistorted North American Craton.

The integrated strengths of $25 \cdot 10^3$ GPa may give an indication of the unstretched oro-
genic lithosphere strength and resembles the estimates of Bardoux and Marschal [1994] of orogenic lithosphere prior to post-orogenic extension. These studies also indicate that such lithosphere would withstand the weight of thickened crust, and that post-orogenic extension may not be expected for lithosphere with primordial heat flow below 60 mW/m², unless the deformation triggered by enhanced mantle activity (delamination or small-scale convection). The paleo-rheological structure immediately after the orogeny is indeed relevant for understanding the geodynamics behind the development of the discontinuity. In addition, not only the formation of, but the presence of the discontinuity itself may also be relevant for understanding the FFB and foreland basin thermal and uplift history.