The role of partial melting and extensional strain rates in the development of metamorphic core complexes

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Orogenic collapse involves extension and thinning of thick and hot (partially molten) crust, leading to the formation of metamorphic core complexes (MCC) that are commonly cored by migmatite domes. Two-dimensional thermo-mechanical Ellips models evaluate the parameters that likely control the formation and evolution of MCC: the nature and geometry of the heterogeneity that localizes MCC, the presence/absence of a partially molten layer in the lower crust, and the rate of extension. When the localizing heterogeneity is a normal fault in the upper crust, the migmatite core remains in the footwall of the fault, resulting in an asymmetric MCC; if the localizing heterogeneity is point like region within the upper crust, the MCC remains symmetric throughout its development. Therefore, asymmetrically located migmatite domes likely reflect the dip of the original normal fault system that generated the MCC. Modeling of a severe viscosity drop owing to the presence of a partially molten layer, compared to a crust with no melt, demonstrates that the presence of melt slightly enhances upward advection of material and heat. Our experiments show that, when associated with boundary-driven extension, far-field horizontal extension provides space for the domes. Therefore, the buoyancy of migmatite cores contributes little to the outer envelope of metamorphic core complexes, although it may play a significant role in the internal dynamics of the partially molten layer. The presence of melt also favors heterogeneous bulk pure shear of the dome as opposed to the bulk simple shear, which dominates in melt-absent experiments. Melt presence affects the shape of P-T-t paths only slightly for material located near the top of the low-viscosity layer but leads to more complex flow paths for material inside the layer. The effect of extension rate is significant: at high extension rate (cm yr−1 in the core complex region), partially molten crust crystallizes and cools along a high geothermal gradient (35 to 65 °C km−1); material remains partially molten in the dome during ascent. At low strain rate (mm yr−1 in the core complex region), the partially molten crust crystallizes at high pressure; this material is subsequently deformed in the solid-state along a cooler geothermal gradient (20 to 35 °C km−1) during ascent. Therefore, the models predict distinct crystallization versus exhumation histories of migmatite cores as a function of extensional strain rates. The Shuswap metamorphic core complex (British Columbia, Canada) exemplifies a metamorphic core complex in which an asymmetric, detachment-controlled migmatite dome records rapid exhumation and cooling likely related to faster rates of extension. In contrast the Ruby Mountain-East Humboldt Ranges (Nevada, U.S.A.) exhibits characteristics associated with slower metamorphic core complexes.

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1. Introduction

Metamorphic core complexes (MCC) are common crustal-scale features that develop when thermally mature orogens undergo horizontal extension. This extension can be driven either by volume forces (active extension) or surface forces (passive extension) following a change in plate kinematics. Migmatite-cored MCC develop when partially molten continental crust is exhumed beneath low-angle normal faults (detachments) to form the core of high-grade structural domes. McMCC offer an opportunity to study deep crustal processes; they contribute to the thermal and mechanical re-equilibration and differentiation of orogenic crust, and, as we show here, record the kinematic boundary conditions prevailing during the final stages of orogeny. The structural and thermal characteristics of McMCC result from the interplay of plate boundary forces, viscous forces (strength of the lithosphere) and body forces (induced by lateral variation in densities), and from syn-deformational processes that affect the rheology and/or buoyancy of the system, such as partial melting and strain localization. Physical and numerical experiments have shown that strain localization is essential to generate
metamorphic core complexes and is easily achieved around rheological and/or density anomalies such as a pre-existing detachment fault in the brittle upper crust (Buck, 1993; Lavier et al., 1999; Koyi and Skelton, 2001; Dyksterhuis et al., 2007; Gessner et al., 2007), a rheological and/or density anomaly in the lower crust (Brun et al., 1994; Brun, 1999; Teyssier et al., 2004, 2008), or a severe density jump along the brittle-ductile transition (Wijns et al., 2005). More recently, Regenauer-Lieb et al. (2006) demonstrated that localization takes place by natural physical processes even in the absence of any particular weakness zone. The shape and orientation of the strain localization agent may have important consequences for the asymmetry of McMCC. In particular, when strain localization is achieved via a weak layer beneath the brittle/ductile transition, normal fault bounded symmetric graben accommodate extension in the upper crust (Teyssier et al., 2004, 2008). In contrast, asymmetric flow unfolds when a detachment-like rheological anomaly occurs in the upper crust (Lavier et al., 1999). This detachment fault rotates into a subhorizontal décollement (a rolling-hinge detachment) with ongoing shear strain localized along the hanging wall block (Brun et al., 1994; Lavier et al., 1999). Upon extension, localized thinning of the upper crust is isostatically compensated by the flow of lower crust into the MCC (Block and Rendu, 1990; Wdowinski and Axen, 1992). A weak lower crust is therefore important for metamorphic domes to grow (Brun et al., 1994; Wijns et al., 2005; Teyssier et al., 2008). The viscosity of partially molten rocks decreases dramatically with increasing melt fraction (Richet and Bottinga, 1995; Scaillet et al., 1996; Baker, 1998, Arbaret et al., 2007) and, in the common case of fluid-absent partial melting, density also decreases (Clemens and Droop, 1998). Because most fluid-absent melting reactions have positive dP/dT slopes, a positive feedback between extension induced decompression and melting could be important in the structural and thermal development of McMCC (Teyssier and Whitney, 2002; Whitney et al., 2004). The degree of coupling between the brittle upper crust and the ductile lower crust is also known to impact the structural evolution of MCC. Physical experiments show that a strongly coupled continental crust leads to distributed surface extension, whereas a mechanical decoupling between a very weak lower crust and a much stronger upper crust leads to localized surface extension, whereas a mechanical decoupling between a very weak lower crust and a much stronger upper crust leads to localized surface extension (Brun et al., 1994; Brun, 1999). Partial melting produces a strong decoupling because it reduces drastically the viscosity of the lower continental crust. Very few studies have investigated the role of partial melting in the development of McMCC (Teyssier et al., 2004, 2008) and none of them take into account temperature-dependent melt fraction and/or melt-dependent viscosity and density. We have therefore investigated the thermal and mechanical evolution of McMCC via a series of 2D numerical experiments that address the impact of partial melting on temperature, density, and viscosity. In this paper, we compare simulation results for cases involving a partially molten (low viscosity) vs solid-state lower crust for different extension rates and different types of strain localization agents. Strain, flow paths, and P-T-T paths generated from the models allow us to better interpret first-order characteristics of McMCC.

2. Numerical code and model setup

We use Ellipsis, a Lagrangian integration point finite-element code, to solve the governing equations of momentum, mass and energy in incompressible flow (Moresi et al., 2001, 2002, 2003). The equations of motion are solved using a multigrid solver coupled to a continuum equation via a multigrid Uzawa iteration. The explicit time-stepping of a Petro-Galerkin upwind method solves the energy equation (Moresi and Galeri, 1995, 1998). In Ellipsis, the stress tensor is decomposed into pressure \( p \) and deviatoric stress \( \tau \):

\[
\sigma_{ij} = \tau_{ij} - p \cdot \delta_{ij}
\]  

subject to the incompressibility condition: \( \nabla \cdot v = 0 \), in which \( v \) is the velocity. Acceleration is neglected as material deforms via slow viscous flow. Temperature derives from solving the energy equation:

\[
\frac{DT}{Dt} = \frac{\partial T}{\partial t} + v \cdot \nabla T = \kappa \Delta^2 T + H
\]  

with \( \kappa \) the material time derivative, \( T \) the temperature, \( \kappa \) the thermal diffusivity and \( H \) the rate of energy production (in our models this amounts to the radiogenic heat in the continental crust and the constant heat flow at the base of the models). Ellipsis has been used to model crustal deformation in extension (Wijns et al., 2005; Gessner et al., 2007) and compression (Mühlhaus et al., 2002). We have implemented a melt function to account for the thermal and mechanical effects of partial melting (cf. O’Neill et al., 2006 for details). This function however doesn’t account for melt extraction processes (i.e. compaction or diking) so the code is restricted to models involving partially melted regions in which the melt remains in its source. For the melt fraction \( F \), Ellipsis uses McKenzie and Bickle (1988) parameterization, which expresses \( F \) as a function of the super-solids temperature \( T_{ss} \), the solidus temperature \( T_{sol} \) and the liquidus temperature \( T_{liq} \):

\[
T_{ss} = \frac{T_{sol}(\alpha) + T_{liq}(\alpha)}{T_{liq}(\alpha) - T_{sol}(\alpha)}
\]

The melt fraction is then:

\[
F(T_{ss}) = 0.5 + T_{ss} \cdot \left( T_{ss}^2 - 0.25 \right) \cdot (0.4256 + 2.988 \cdot T_{ss})
\]

The density of partially melted regions changes with both \( T \) and \( F \):

\[
\rho(T,F) = \rho_0 \cdot (1 - \alpha \cdot \Delta T - F \cdot \beta)
\]

where \( \alpha \) is the coefficient of thermal expansion and \( \beta \) is the coefficient of expansion related to phase change. The starting model (Fig. 1) is 270 km long and 90 km thick, and includes 15 km of air-like material (low density, low viscosity) and a 60 km thick crust above 15 km of upper mantle. A weak prismatic region dipping 45° simulates a detachment fault in the upper crust. The geotherm is based on a constant heat flow imposed at the base of the model (0.022 W m\(^{-2}\)), a constant temperature imposed at the top (20°C), a depth independent radiogenic heat production in the crust (767·10\(^{-7}\) W m\(^{-2}\)) and zero heat flow across the lateral sides of the model. For all rocks, we assume a heat capacity of 1000 J kg\(^{-1}\) K\(^{-1}\) and thermal diffusivity of 9·10\(^{-7}\) m\(^2\) s\(^{-1}\). The crustal thermal conductivity is therefore 2.45 W m\(^{-1}\) K\(^{-1}\). The temperature field is allowed to equilibrate until the Moho of the 60 km thick crust reaches 975°C.

Because the composition of the continental crust changes with depth, an increase in density is expected toward the Moho. However, we assume that this density increase is balanced by the density decrease related to thermal expansion, and therefore we choose a depth and temperature independent density for the continental crust of 2720 kg m\(^{-3}\). At room conditions, the density of the mantle is 3370 kg m\(^{-3}\) and has a coefficient of thermal expansion of 2.8·10\(^{-5}\) K\(^{-1}\).

The crust and the mantle have a visco-plastic rheology with a temperature and stress dependent viscosity for stresses below the yield stress, and a depth dependent plastic branch above it. We use power law relationships between strain rate and stress to describe dislocation creep. The viscosity varies with the temperature and stress according to:

\[
\eta = \frac{1}{2} \cdot \frac{A}{Q} \cdot \exp \left[ \frac{Q}{RT} \right] \cdot \varepsilon^{(1-n)/n}.
\]  

For the continental
Fig. 1. Model geometry, parameters and boundary conditions, showing the solidus and liquidus, temperature, melt fraction, strength as a function of depth, and the locations of Lagrangian markers used to track temperature and pressure history, flow paths and finite strain in the central part of the model. The density in crust decreases with the melt fraction (cf. text) but is otherwise temperature independent. The models are 270 km long and 90 km deep. They include 15 km of air-like material, a 60 km thick continental crust and 15 km of mantle. A weak detachment-like rheological anomaly is introduced in the upper brittle crust to localize deformation. The thermal parameters are such that a 40 km thick crust would have a steady-state Moho temperature at 540 °C. Following thickening, thermal relaxation was let to proceed until the Moho temperature reached 975 °C. Then, extension is driven at both ends of the model by imposing a constant velocity condition.

Fig. 2. Comparison of experiments in which strain localization is initiated and proceeds from a point-like rheological anomaly located at the brittle/ductile transition (left column), and the same experiment (right column) with a detachment-like rheological anomaly dipping 45° to the right and cutting through the entire brittle crust and slightly extending into the ductile crust. In both experiments, strain rate is $2 \times 10^{-15}$ s$^{-1}$. 
crust, $A_n$ and $Q$ are those of quartz-rich rock (pre-exponent $A_n = 5 \cdot 10^{-6}$ MPa$^{-n_c}$ s$^{-1}$, $n_c = 3$, $Q_c = 1.9 \cdot 10^5$ J mol$^{-1}$, Goetze, 1978). The pre-exponent factor for the fault is $10 \cdot A_n$, which makes the fault ten times less viscous than the crust in which it is embedded. The rheology of the mantle is that of dry olivine (pre-exponent $A_m = 7 \cdot 10^4$ MPa$^{-n_m}$ s$^{-1}$, $n_m = 3$, $Q_m = 5.2 \cdot 10^5$ J mol$^{-1}$, Brace and Kohlstedt, 1980).

In the crust and the mantle, frictional sliding is modeled via a Mohr Coulomb criterion with a cohesion ($C_0$) of 15 MPa and a coefficient of friction ($\mu$) of 0.44. The cohesion and coefficient of internal friction in the detachment fault are $C_0/10$ and $\mu/10$, respectively. In all material, the yield stress linearly drops to a maximum of 20% of its initial value when the accumulated strain reaches 0.5 (cf. Wijns et al., 2005 for details). For differential stresses reaching the yield stress, the material fails and deformation is modeled by an effective viscosity: $\eta_{\text{yield}} = \tau_{\text{yield}}/(2 \cdot E)$ in which $E$ is the second invariant of the strain rate tensor. For semi-brittle effects, we impose a maximum yield stress of 250 MPa for the crust, 400 MPa for the mantle and 10 MPa for the fault.

The crust has a solidus and liquidus with positive $dP/dT$ that is representative of fluid-absent partial melting reactions (Clemens and Droop, 1998). We did not attempt to use the solidus/liquidus of any specific felsic lithology. The solidus and liquidus, which are defined as polynomial functions of temperature, are adjusted to obtain a peak melt fraction of 35% at the Moho (Fig. 1). A latent heat of fusion of $250$ kJ kg$^{-1}$ K$^{-1}$ is embedded into the energy equation. The density of the partially melted region decreases linearly by 13% between the solidus and the liquidus (Clemens and Droop, 1998). In nature, the viscosity of partially melted regions likely drops over many orders of magnitude when the melt forms a connected network (Arzi, 1978; Van der Molen and Paterson, 1979; Scaillet et al., 1996). In our experiments, the viscosity decreases linearly by 3 orders of magnitude when the melt fraction increases from 15 to 30%. When the melt fraction is < 15%, the viscosity of the melted crust is that of the non-melted surrounding; when the melt fraction is > 30%, its viscosity is a thousand times lower than in surrounding material. Rosenberg and Handy (2005) showed that significant weakening occurs at 7% melt fraction. In our experiments, the melt fraction increases up to 30% over a few kilometres. Consequently, the results are not affected by shifting this critical melt fraction down to 2–12%. In the models, there is no segregation of the melt from its source. Although this is a limitation of the code, it is a reasonable approximation for many McMCC in which melt and solid fractions move en masse (Teyssier and Whitney, 2002; Whitney et al., 2004).

Moving boundary conditions are imposed at both lateral ends of the model. We have tested a range of initial strain rates in between $6 \cdot 10^{-15}$ s$^{-1}$ and $2 \cdot 10^{-16}$ s$^{-1}$ (i.e., 25.5 mm/y to 0.85 mm/y at both sides). These strain rates are only a guide, as they are achieved for the first increment of deformation and decrease as extension proceeds.

Fig. 3. All models have recorded 25% extension following 40 m.y. of extension at a strain rate of $2 \cdot 10^{-16}$ s$^{-1}$ (a, c), and 4 m.y. of extension at a strain rate of $2 \cdot 10^{-15}$ s$^{-1}$ (b, d). (a, b) Models with melt function turned off, contoured for temperature. (c–d) Models with melt function turned on contoured for temperature (c1–d1) and melt fraction (c2–d2). Flow paths of the deeper grid nodes are shown as thick white lines.
Since deformation is strongly localized around the detachment fault, much higher strain rates are achieved around the MCC. A free slip condition is imposed on both the upper and lower horizontal boundaries of the models, and no vertical motion is possible along the base of the models. The latter condition could limit somehow the capacity for the Moho to rise during extension. However, the hot geotherm, which makes the sub-continental mantle very weak, enhances the ability of the mantle to flow horizontally and vertically above the base of the model, hence allowing the Moho to move vertically.

3. MCC geometry, temperature and partial melt distribution

The distribution and nature of extensional features at the surface are strongly affected by the distribution and shape of rheological anomalies that existed in the crust before the onset of extension (e.g. Dyksterhuis et al., 2007). A point-like anomaly results in symmetric extension accommodated by a graben bounded by two normal faults (Brun et al., 1999 and references therein) (Fig. 2). As extension proceeds \( (2 \cdot 10^{-15} \text{ s}^{-1}) \) to 25% surface extension, both normal faults remain active. A migmatite core, which crystallizes during its exhumation, is exhumed at the center of the graben, which remains symmetric. In contrast, a normal fault anomaly cutting through the upper crust leads to asymmetric extension accommodated by a rolling-hinge detachment (Brun, 1994; Lavier et al., 1999). As extension proceeds, strain localizes along the margin of the hanging wall, and the migmatite core is asymmetrically exhumed beneath the active segment of the detachment fault. Despite the symmetry of kinematic boundary conditions, the crystallized migmatite core is systematically shifted toward the hanging wall. This shift does not exist when a point-like rheological anomaly is the strain localization agent. Therefore, the shift of the dome is most likely due to the initial dip direction of the planar zone of weakness. In nature, the position of migmatite cores relative to MCC detachments could be used to infer the dip direction of the original detachment fault.

Models deformed at relatively low \( (2 \cdot 10^{-16} \text{ s}^{-1}) \) and high \( (2 \cdot 10^{-15} \text{ s}^{-1}) \) extension rates are compared for the same surface extension (25%) reached after 40 m.y. (lower) and 4 m.y. (higher) extension rates (Fig. 3). In melt-absent experiments (Fig. 3a and b), higher extension rate promotes significant heat advection (relief of isotherms, Fig. 3b) and internal deformation of material in the upwelling region, resulting in a slightly wider dome. This result seems at odds with results from physical experiments, which suggest that fast extension leads to distributed surface extension (Brun et al., 1994; Brun, 1999). This discrepancy is likely due to the more realistic visco-plastic stratification in our numerical experiments compared to the shear stress discontinuity at the sand/silicone contact in physical experiments.

With the melt function turned on (Fig. 3c1–2 and d1–2), the core complex broadens slightly relative to the no-melt cases. Slow extension with melt (Fig. 3c1–2) enhances upward motion of deep crust compared to the no-melt case (Fig. 3a), resulting in larger exhumation of rocks that were once partially molten. Fig. 3c2 shows an antiform of crystallized partial melt (in blue), akin to a 3D migmatite dome that represents preferential exhumation of deep rocks in the detachment footwall. The solidus at 40 m.y. (Fig. 3c2) is relatively flat and located at about 27 km depth. In contrast, fast extension (Fig. 3d1–d2) produces greater heat advection that results in upward motion of the partial melt layer. After 4 m.y., the solidus is located at ca. 7 km depth and the partial melt antiform is cored by a region of >30% melt at shallow levels (Fig. 3d1). The exhumation magnitude of the solidus, and the crystallization versus exhumation/deformation history, are the most fundamental differences between models extending at fast and slow strain rates. Initially at a depth of 37 km, fast extension brings the solidus to about 7 km to the surface in
a few m.y. In this case, cooling and crystallization of partially molten crust occurs after the bulk of deformation accommodating the exhumation of the dome, possibly even after extension has stopped, as the solidus recedes downward during cooling of the crust. In contrast, slow extension brings slowly the solidus to about 27 km to the surface over a few tens of my. In this case, a large volume of migmatite is advected upward through a quasi-static solidus (blue region in Fig. 3c), as rocks are exhumed faster than the solidus. In the latter case, the bulk of deformation accommodating the dome exhumation occurs after crystallization, and partial melting is restricted to $P > 600$ MPa ($\geq$ ca. 22 km). This contrasted crystallization versus exhumation/deformation history suggests that solid-state deformation is dominant in slow extension MCC, whereas weakly deformed domes are expected in fast extension MCC as crystallization and cooling post-date the bulk of exhumation of the dome. The contrasting crystallization versus exhumation/deformation history also suggests that low-pressure migmatites ($P < 400$ MPa, i.e. < ca. 15 km) require fast extensional strain rates, whereas migmatites restricted to medium to high-pressure ($P > 600$ MPa) are prevalent in slow extension MCC.

4. Finite strain and $P$-$T$-$t$ paths

Although the various MCC in Fig. 3 have similar shapes, their internal deformation differs significantly. This is documented in Fig. 3 by contrasted flow paths and contrasted finite strains recorded by a passive grid. In experiments with no melt, the bulk of the MCC region is involved in a combination of counter-clockwise rotation (dominant in Fig. 3a), top to the left heterogeneous simple shear, and pure shear (dominant in Fig. 3b). Counter-clockwise rotation is the consequence of...
of the interplay between asymmetric simple shear and isostasy, which produces a horizontal exhumation gradient, as shown by the flow paths in Fig. 3a. The top right corner of the grid is caught in the detachment zone, recording a top to the right shearing. Prominent early upward motions and later minor horizontal motion characterize the flow path of the lowermost grid nodes (Fig. 3a and b). In melt present experiments, the later horizontal motion is more developed and heterogeneous bulk flattening dominates the finite strain of the grid, except in the vicinity of the active segment of the detachment fault, where top to the right simple shear dominates. Away from the active segment of the detachment, there is a strong contrast between the bulk simple shear that dominates the finite strain in melt-absent fast extension experiments, and the bulk pure shear which dominates in slow extension melt-present experiments.

\( P-T-t \) paths are very sensitive to strain rate but are not affected by the presence or absence of melt (Fig. 4), as long as convection does not affect the particles. At low extensional strain rate (Fig. 4a and c), \( P-T-t \) paths exhibit decompression and cooling along a geothermal gradient in the range of 20 to 35 °C km\(^{-1}\), with similar shapes in both melt and no-melt experiments. In all experiments, the 20 km marker, located structurally above the detachment fault (i.e., in the hanging wall), records near isobaric heating. This heating is due to the advection of hot material that occurs preferentially beneath the hanging wall. At higher strain rate (Fig. 4b and d), most \( P-T-t \) paths show an episode of near isothermal decompression followed by a period of both decompression and cooling along a hotter geothermal gradient (35–65 °C km\(^{-1}\)). Deep markers have \( P-T-t \) paths crossing from the kyanite to the sillimanite stability field, whereas shallower markers cross from the kyanite to the andalusite stability field. At high strain rate, and when melt is present, some \( P-T-t \) paths show a sudden change from near isothermal decompression to near isobaric cooling (Fig. 4d) consistent with the observed change in direction of the particle flow paths.

5. Effects of critical melt fraction and buoyancy

The flow paths of particles in fast McMCC models (Fig. 5a) are affected not only by the nature of the rheologic heterogeneity that localizes emplacement of the MCC (point vs. fault, Figs. 2 and 5b), but also by the fraction of melt in the deep crust and the associated variations in viscosity and density structure. In the experiments involving the presence of melt (Figs. 2 and 3c and d), viscosity was dropped between 15 and 30 melt %, but what is the effect of lowering this melt fraction on the behavior of the McMCC? A viscosity reduction of three orders of magnitude between 2 and 12 melt % results in weakening this melt fraction on the behavior of the McMCC? A viscosity reduction of three orders of magnitude between 2 and 12 melt % results in weakening of the partially melted region at shallower depth, and therefore over a larger volume (Fig. 5c). However, this effect is not sufficient to modify significantly the large-scale behavior of McMCC (cf. Fig. 5a and c). In the model initial condition (Fig. 1), the melt fraction increases nonlinearly (McKenzie and Bickle, 1988), reaching 30% melt at about 8 km below the solidus. Therefore, a decrease in the critical melt fraction, at which a large viscosity reduction occurs, has no significant rheologic effect. A low critical melt fraction may have significant implications for melt migration and local rheology (e.g., Rosenberg and Handy, 2005) that are not investigated here, but have little impact on the large-scale development of McMCC in both fast extension and slow extension cases.

Another important question concerns the role of buoyancy in the upward flow of a partially molten layer. In order to test the buoyancy effect, we conducted an experiment in which melting was not
accompanied by density decrease in the post-solidus region. This experiment (Fig. 5d) confirms that the buoyancy force produces only minor differences in the geometry of flow in the partial melt layer (Brun et al., 1994; Tirel et al., 2004; Wijns et al., 2005; Gessner et al., 2007; Tirel et al., 2008). The main differences (cf. Fig. 5a and d) are a slight ballooning effect of the partial melt layer and a more continuous channel of high melt fraction above the Moho, and this likely allows a more effective material transfer from the base of the crust into the MCC. However, upward flow is determined mainly by a dynamic feedback between extension of the upper crust and the necessity for the low viscosity lower crust to fill the zone of extension efficiently (Wdowinski and Axen, 1992). In other terms, isostasy rather than buoyancy drives exhumation. This conclusion is valid only when surface extension is the product of kinematic boundary conditions imposed on the lateral sides of the models. Contrasting results can be expected when volume forces, due to lateral variation in gravitational energy, drive extension. In such a case, space for MCC is provided by internal redistribution of masses (Rey et al., 2001), in which case the buoyancy of the melted lower crust is of fundamental importance.

6. The Shuswap (British Columbia, Canada) and Ruby Mountain–East Humboldt Range (Nevada, U.S.A) as examples of faster and slower McMCC

Metamorphic core complexes in the North American Cordillera (Fig. 6a) share first-order characteristics (e.g. detachment faults separating high-grade metamorphic rocks from upper crust), but in detail display a wide range of styles that could be linked to the tempo and duration of extensional deformation, as well as the volume of melt involved. The formation of most of the MCC in the northern Cordillera (Nevada, USA to British Columbia, Canada) involved crustal melting and the formation of migmatite domes (McMCC), whereas MCC in Arizona and southern California developed in a primarily solid-state crust and are dominated by a simple shear detachment zone.

The Ruby Mountain–East Humboldt Range (R–EH) McMCC in the Basin and Range (Nevada) and the Shuswap McMCC in the Canadian Cordillera exemplify some of the contrasting features developing under different conditions of melt and extension rate. The geometry, P–T evolution, and extension/exhumation rates of the Shuswap core complex, including the Thor-Odin migmatite dome (Fig. 6b) closely match our models of fast extension in the presence of melt (Fig. 3d). In this migmatite-cored dome, crystallization of leucogranite and migmatite leucosome occurred between 60 and 52 Ma (Vanderhaeghe et al., 1999; Hinchee et al., 2006), followed by rapid cooling through mica $^{40}$Ar/$^{39}$Ar closure temperatures at 49–48 Ma (Vanderhaeghe et al., 2003; Mulch et al., 2004; Teyssier et al., 2005). The Thor-Odin migmatite core is shifted toward the eastern detachment (Columbia River), suggesting that this fault zone acted as an E-dipping rolling-hinge that separated the cool and thick foreland crust to the east from the partially molten crust in the hinterland (Teyssier et al., 2005). The P–T–t path of dome rocks records near-isothermal...
decompression from 800–1000 MPa to ~500 MPa at ca. 750 °C (Fig. 6c; Norlender et al., 2002) under partial-melt conditions, indicating that melt crystallized at low pressure as in our models (cf. Fig. 3d). The Thor-Odin example is representative of McMCC in the northern Cordillera, as well as other core complexes that contain migmatite domes (e.g., Aegean; Himalaya).

In contrast, the Ruby Mountain–East Humboldt Range McMCC (Fig. 7a) records a protracted history of extension and exhumation that may have started as early as Late Cretaceous and continued during the Eocene to the mid-to late Oligocene, when the bulk of McMCC development occurred (e.g. McGrew et al., 2000; Sullivan and Snake, 2007). Partial melting in the lower crust, synchronous with extension and exhumation, is evidenced by late Eocene and mid Oligocene granitic plutons, dikes and sills strongly transposed into the extensional mylonitic fabric (Wright and Snake, 1993) (Fig. 7b). In the migmatite dome, metamorphic assemblages record, from ca. 85 Ma to ca. 55 Ma, a first episode of syn-depositional melting and cooling from >900 MPa and 800 °C to 700 MPa and 650 °C, along a path sub-parallel to the kyanite-sillimanite transition (McGrew et al., 2000) compatible with slow exhumation and crystallization at depth. From ca. 55 to ca. 40 Ma, decompression developed under near constant temperature from 700 MPa and 650 °C to 300 MPa and 600 °C (Fig. 7c). According to our experiments, this second episode is compatible with faster exhumation. Finally between 35 and 22 Ma, the third episode of extension, the main MCC phase, involved coeval decompression and cooling along a temperature gradient of 40 °C km⁻¹ (McGrew et al., 2000), which, according to our experiments, suggests a slower extension rate.

7. Conclusions

Numerical experiments on the development of MCC reveal first-order petrological and structural features that can be identified in the field to gain insights into the geometrical and thermal conditions prevailing before, and kinematic conditions prevailing during, the development of MCC. These experiments reveal fundamental differences between two end-member classes of migmatite-cored (Mc) MCC. In one case, McMCC developed under fast localized extension (on the order of cm/yr) with weakly deformed migmatite cores that record near-isothermal decompression and partial melting before cooling along a high geothermal gradient (ca. 35 to 65 °C km⁻¹) and crystallizing at low pressure. In the other case, McMCC that develop during slower local extension (on the order of mm/yr) have migmatite cores that crystallize at high pressure (~600 MPa) before accumulating solid-state deformation during coeval decompression and cooling along a lower thermal gradient (ca. 20 to 35 °C km⁻¹). A predicted shift of the migmatite core towards the hanging wall is due to strain localization imposed by a detachment fault (or any dipping rheological heterogeneity), which in turn imposes an asymmetric flow field and therefore an asymmetric strain pattern. From this shift, the original dip of the detachment fault can be inferred. Another important result is that the rheologically critical melt fraction at which the viscosity drops by several orders of magnitude as a function of melt content plays only a minor role in the crustal-scale development of McMCC; further, the buoyancy force related to the presence of less dense melt in the partial melt crust has little influence on the geometry and evolution of McMCC. The northern Cordilleran metamorphic core complexes in the northern US and British Columbia, Canada, exemplify faster McMCC in which the collapse of thick/hot crust and the formation of metamorphic core complexes took only a few million years. These McMCC display a series of migmatite domes that are located on one side of the core complex and in the immediate footwall of a rolling-hinge detachment zone, suggesting that the detachment zone controlled the asymmetric development of the migmatite domes. The core migmatites underwent near-isothermal decompression followed by rapid cooling over only a few million years and fit the first-order features of material involved in fast extension models. Although the Ruby Mountain–East Humboldt Range exhibits an episodic extension history over a relatively long period of time, it fits in first approximation the thermal and structural characteristics of slower McMCC.

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References


