Crustal structure: A key constraint on the mechanism of ultra-high-pressure rock exhumation

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A B S T R A C T

The distribution of ultra-high-pressure (UHP) metamorphic rocks demonstrates that burial (to >100 km) and rapid exhumation (>1 cm a−1) of continental crust is a normal part of early (~10 Ma) continental collision. Currently, there is no comprehensive model for this fundamental tectonic process that also satisfactorily explains the upper-crustal structures resulting from early collisional UHP rock exhumation. Characteristic features requiring explanation include: structural domes that are cored by UHP nappes; associated medium- to high-pressure nappes displaying a distinct “pressure gap”; overlying lower-grade rocks, including suture zone ophiolites; and, coeval foreland-directed thrust-faults and syn-exhumation normal faults. We present a geometrical model involving crustal burial and exhumation in a subduction channel below an accretionary wedge. Competition between down-channel shear traction and up-channel buoyancy forces, expressed as the exhumation number, E, controls burial and exhumation, leading to rapid up-channel flow when E > 1. Exhuming UHP material forms a nappe stack and structural dome as it penetrates and destabilises the overlying wedge, driving thrusting and extension. This solution is compelling because it explains both the geology and the petrology of the Tso Morari and other UHP complexes, and because it demonstrates that pulse-like buoyant exhumation from deep in the subduction channel creates observed upper crustal structures. This places constraints on the exhumation mechanism and provides a test of alternative models. Other proposed mechanisms, such as continuous circulation in a lithospheric-scale wedge or overpressured subduction channel, predict different types of upper-crustal structures and are therefore unsatisfactory explanations for early collisional exhumation of UHP terranes.

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1. Introduction

Twenty-five years ago, the discovery of coesite in crustal rocks (Chopin, 1984; Smith, 1984) led to the recognition of a new category of ultra-high-pressure (UHP) metamorphism and ultimately to widespread acceptance that continental subduction is a normal part of early collisional orogenesis (Ernst et al., 1997; Chopin, 2003). Since then, a wealth of field and analytical data (e.g., Dobrzhinetskaya et al., 1995; Ernst et al., 1997; Chopin, 2003; Rubatto and Herrmann, 2001; de Sigoyer et al., 2004; Terry and Robinson, 2004; Jolivet et al., 2005; Hacker, 2006; Johnston et al., 2007; Epard and Steck, 2008) and physical and numerical experiments (e.g., Chemenda et al., 1995, 2000; Roselle and Engi, 2002; Gerya et al., 2002, 2008; Yamato et al., 2008) have offered insight into the origin and evolution of UHP terranes. However, beyond general agreement that the buoyancy of subducted crust contributes to its exhumation (England and Holland, 1979; Platt, 1993; Ernst et al., 1997), a comprehensive model accurately linking subduction channel processes to observed upper-crustal geology remains elusive.

The defining petrological characteristics of UHP terranes are well known. Most record peak metamorphic conditions in the range 2.6–4.0 GPa and 600–800 °C (e.g., Chopin, 2003; Hacker, 2006), with significantly higher peak PT conditions reported from some locations (e.g., Kaneko et al., 2000; Fauré et al., 2003). In many cases, maximum burial was followed by rapid exhumation at 1–3 cm a−1 from depths exceeding 100 km to lower crustal levels (e.g., Rubatto and Herrmann, 2001; Parrish et al., 2006), with the whole process completed within ca. 10 Ma of initial collision (referred to here as “early collisional”). Crustal protoliths are typically continental margin sedimentary and crystalline rocks, with diagnostic coesite- and microdiamond-bearing assemblages present in mafic enclaves hosted by metasedimentary and granitoid host gneisses that rarely preserve UHP minerals. In addition, the upper-crustal settings of many UHP terranes require a number of structural characteristics that should be accounted for by any viable tectonic model (Fig. 1, numbers 1–6). UHP complexes typically occupy the cores of antiformal nappe stacks that define structural domes (1), ranging from <5 to >50 km across (Avigad et al., 2003; Fauré et al., 2003; Xu et al., 2006), flanked by low-grade, accretionary wedge and/or upper-crustal sedimentary rocks (2). The high-grade domes and adjacent lower-grade units are spatially associated with, and may
directly underlie, ophiolitic rocks that mark the collision zone suture (3) (Platt, 1993). Overlying and underlying nappes (4), separated from UHP rocks by ductile shear zones with mainly thrust-sense kinematics (Kaneko et al., 2000; Avigad et al., 2003; Fauré et al., 2003; de Sigoyer et al., 2004), contain high- to medium-pressure assemblages indicating a substantial pressure gap (e.g., Platt, 1993). Steep structures formed at UHP conditions (e.g., Fauré et al., 2003; Terry and Robinson, 2004; de Sigoyer et al., 2004; Epard and Steck, 2008) are generally strongly overprinted by pervasive, typically shallow-dipping, amphibolite facies fabrics. Doming normally post-dates the fabrics associated with nappe juxtaposition, and may be coeval with foreland-directed thrusting (Kaneko et al., 2000; Avigad et al., 2003; de Sigoyer et al., 2004; Johnston et al., 2007).

The presence of these upper-crustal structures in many different UHP terranes, particularly those formed and exhumed during the early stages of collision, suggests that these features may be intrinsic to early collisional UHP exhumation. If so, how do they form? What is the link between deep-seated subduction zone processes and near-surface structures? Any successful model for UHP generation and exhumation must not only exhume UHP material with appropriate PT conditions, have not yet met the second requirement (e.g., Chemenda et al., 1995, 2000; Gerya et al., 2002, 2008; Yamato et al., 2008).

This paper presents a geodynamical model that explains both the mechanics and the geology of the subduction, metamorphism, exhumation, and emplacement of UHP terranes into the upper level of the crust. Our focus is on UHP metamorphism and exhumation that occurs within ca. 10 Ma of the transition from oceanic subduction to continental collision (early collisional) rather than on late-orogenic or multiple UHP exhumation events, where appropriate boundary conditions are less clear. We use a numerical model with an embedded, high-resolution (nested) grid to produce results that can be realistically compared with data from natural UHP terranes. Unlike conceptual or kinematic models, this “working” model is internally consistent and compatible with the physics of the model system. Contrasting flow modes in model subduction channels (Warren et al., 2008a) reflect competition between down-channel shear traction and up-channel buoyancy forces (Ernst et al., 1997), quantified in terms of the exhumation number, $E$ (Raimbourg et al., 2007; Warren et al., 2008a,b). Increasing $E$ during subduction, rather than buoyancy alone, optimises rapid exhumation. Interaction of a rising UHP plume with an overlying accretionary wedge accounts for the near-surface features of UHP terranes (Fig. 1), including a broad spectrum of observations from the Himalayan Tso Morari complex (de Sigoyer et al., 2004; Epard and Steck, 2008).

2. Flow modes in subduction channels

Numerical models of subduction channels are conveniently interpreted in terms of the characteristic exhumation number, $E$ (Raimbourg et al., 2007; Warren et al., 2008a), and corresponding flow modes associated with burial and exhumation of UHP rocks (Fig. 2). The first-order dynamics (Fig. 2) can be approximated by lubrication theory for creeping flows, and characterised in terms of the competition between down-channel Couette flow, caused by the drag of the subducting lithosphere, and the opposing up-channel Poiseuille flow, driven by the buoyancy of low-density subducted crustal material (A1.1). This competition is expressed through the exhumation number $E = \frac{h^2 \Delta P_{\text{eff}}/\partial x}{(n_{\text{eff}} U)}$, a force ratio derived from the non-dimensional channel flow equation (A1.1). Here $\Delta P_{\text{eff}}/\partial x$ is the effective

![Fig. 1. General characteristics of ultra-high-pressure (UHP) complexes and surrounding upper crust (Kaneko et al., 2000; Fauré et al., 2003; Avigad et al., 2003; Xu et al., 2006; Epard and Steck, 2008).](image1)

![Fig. 2. Definition of exhumation number, $E$ (Raimbourg et al., 2007; Warren et al., 2008a), and corresponding subduction channel flow modes (Warren et al., 2008a,b).](image2)
down-channel pressure gradient (x measured in the down-dip direction), \( U \) is the subduction velocity of the lower plate, and \( \eta_{\text{ef}} \) is the effective viscosity of the channel material. The parameter \( h^* = \frac{x}{\eta_{\text{ef}}} \left( \frac{\partial P_{\text{ef}}}{\partial x} \right)^{1/2} \) is a measure of the scale of the thickness of the channel estimated using characteristic initial values \{square brackets\} of viscosity and effective pressure gradient; the actual values that determine \( E \) will depend on the particular problem and its evolving solution. The first bracketed term of \( E \) measures the Poiseuille flow force per unit length of a 2D subduction channel. For a channel with deformable walls and no tectonic over/under-pressure, \( \partial P_{\text{ef}} / \partial x \approx (\rho m - \rho c) g \) Sin\( y \), where \( y \) is channel dip (Fig. 2). The second bracketed term is the Couette traction per unit channel length.

Along with the characteristic \( E \), defined at the scale of the subduction channel, space-time variations in the channel flow can be interpreted in terms of the local exhumation number \( E(x,t) \) and corresponding flow modes (Warren et al., 2008a,b) (Fig. 2). During continental subduction, \( E(x,t) \) evolves from \( <1 \) during burial, to \( \sim 1 \) during detachment and stagnation in the subduction channel, to \( >1 \) at the onset of and during exhumation. Buoyancy is a necessary, but not sufficient, condition for UHP exhumation. Among other controlling factors (Fig. 2), decreasing viscosity \( (\eta_{\text{ef}}) \) is typically most important for driving \( E \) beyond the exhumation threshold (Warren et al., 2008b). In general, \( E(x,t) \) should be regarded as a measure of local exhumation potential; as shown below, even where the local threshold value is exceeded \( (E>1) \), efficient exhumation may be impeded by contric-tions \( (s/h) \) or high viscosities \( (\eta_{\text{ef}}) \) further up the channel.

3. Numerical model design

The subduction–collision transition problem is solved at an upper-mantle scale using a 2D, vertical-section, thermal-mechanical finite element model (Fullsack, 1995; Warren et al., 2008a,b). An outline of the modeling procedure is given here, with details in Appendix A2. The model comprises 120-km thick pro- and retro-continental lithospheres separated by 90-km thick oceanic lithosphere, with sub-lithospheric mantle extending to 660 km depth (Fig. 3). Continental interiors have 24-km thick upper/mid crust and 12-km thick lower crust. Oceanic crust is 8 km thick. The subducting pro-continent interior is bordered by a continental margin, 450 km wide at the surface (Fig. 3b). Margin upper/mid-crust has higher radioactive heat production \( (A = 2 \mu W m^{-3}) \) than the interior \( (A = 1.15 \mu W m^{-3}) \), to represent younger, weaker, less refractory sediments or terranes bordering older, stronger, more refractory continental crust. Pro-continental lithosphere converges on the stationary retro-lithosphere at rates that vary among the models from \( v_T = 5 \) to 15 cm a\(^{-1}\) (Fig. 3, Table 1) causing subduction of the oceanic lithosphere followed by continent–continent collision (Fig. 3). The upper boundary of the model is a stress-free surface. The sides and base of the sub-lithospheric mantle domain have no-slip and free-slip boundary conditions respectively.

The models differ from those we published previously (Warren et al., 2008a,b) in that calculations are made at two scales to improve model resolution. We have implemented a “nested/embedded” version of the ALE fine-element software (Fullsack, 1995) in which a small-scale (SS) sub-domain is embedded within the standard large-scale (LS) model Eulerian domain (Fig. 3a; A2.1). For each time step the large-scale model solves the problem for the entire 2000×660 km domain. LS results are used to define velocity and temperature boundary conditions on the smaller scale domain containing the subduction channel and adjacent regions. The SS model re-solves the problem at higher resolution on the SS domain \( (2 \times 2 \text{ km fine element mesh vs. } 10 \times 2 \text{ km for the LS domain}) \), and the SS solution replaces the LS one within the nested domain. The two domains share a single cloud of Lagrangian tracking particles.

The velocity field, strain rate, deformation, and stress, subject to specified velocity boundary conditions, are calculated using an Arbitrary Lagrangian Eulerian (ALE) methodology (Fullsack, 1995) which solves large-deformation flows with free upper surfaces on an Eulerian finite-element grid that stretches vertically to conform to the material domain. A Lagrangian grid and passive tracking particles, advected with the model velocity field, are used to update the mechanical and thermal property distributions on the Eulerian grid. The heat balance equation is solved on the LS and SS Eulerian grids subject to the thermal boundary conditions and the mechanical velocity field in the model domain, and includes shear heating converted from rate of work at 100% efficiency.

The model solves the equilibrium force-balance and heat equations, with incompressible flow except during phase transitions (A2.2, A2.5). Thermal and mechanical properties are coupled through the

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**Fig. 3.** Model design and main parameters, reference Model V5. Pro-continental and oceanic lithosphere converge on, and subduct beneath, stationary retro-lithosphere at \( v_T = 5 \text{ cm a}^{-1} \), with subduction initiated at the weak zone (Methods). Nested, small-scale (SS) domain (dashed box) is a high-resolution grid embedded in the standard large-scale (LS) model domain \( (x = 1050–1850 \text{ km}; z = 0–150 \text{ km}) \). Flow laws: WO = wet olivine (Karato and Wu, 1993), WQ = wet quartzite (Gleason and Tullis, 1995), DMD = dry Maryland diabase (Mackwell et al., 1998); \( f = \) viscosity scaling factor; \( W_s = \) strain weakening factor (over strain range 5–10; Warren et al., 2008a,b). Retro-continental thermal–mechanical parameters are identical to pro-continental interior. Vertical arrows show heat flux. Other parameter details in Table 1 and Appendix A2.
<table>
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<th>Lower crust</th>
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These values are held constant for all models presented.  
bᵇ (tensor invariant stress pre-exponential factor) = 2^(1−n/α) × 3^(−1/3×2α) A₀⁻¹/α, where A₀ is the uniaxial pre-exponential factor.
thermal activation of viscous flow, shear heating, redistribution of radioactive crust, temperature- and phase-dependent buoyancy forces, and volume changes associated with material phase changes (A2.5). Material rheologies include both frictional-plastic (brittle) and viscous (ductile) deformation, the latter based on laboratory flow laws (A2.3). Materials undergo limited viscous strain weakening, designed to represent the combined effects of reaction- and strain-weakening mechanisms (e.g., Jolivet et al., 2005; Raimbourg et al., 2007; Warren et al., 2008a,b) (A2.4). Model sensitivity to heat production, convergence velocity, margin width and rheology, material densities, and
other properties was investigated previously (Warren et al., 2008a,b) and these parameters are kept constant here (Table 1). Surface processes are not included. Although erosion clearly plays a role in thinning the crust and in associated late-stage exhumation of UHP complexes once they have reached the middle to upper crust (e.g., Platt, 1993), our results demonstrate that surface erosion per se is not required for exhumation of (U)HP material from mantle to lower crustal depths (e.g., Warren et al., 2008a,b). This is also supported by observations that typically show the primary phase of UHP exhumation to be an order of magnitude faster than surface erosion. (e.g., Platt, 1993; Rubatto and Herrmann, 2001).

Ductile deformation is calculated using power-law viscous flow for which the effective viscosity is:

$$\eta^e = \frac{f}{W_s} \frac{B^*}{J_2} (1 - n/2n) \exp \left( \frac{Q + PV^*}{nRT_k} \right)$$

where $f$ is a viscosity scaling factor, $W_s$ is a strain-weakening factor, $B^*$ is the ex-reactive exponent, converted to the tensor invariant form (Table 1), $J_2$ is the second invariant of the deviatoric strain rate, $n$ is the stress exponent, $Q$ is the activation energy, $P$ is the pressure, $V^*$ is the activation volume for power-law creep, $T_k$ is the absolute gas constant. Initial viscous strength of the continental margin is expressed by WQ, $f = 0.4$, and strain weakening of all crustal materials, expressed by $W_s = 10$, represents the combined effects of strain and reaction weakening. This choice and those for other materials are explained in A2.3.

4. Reference model results

Results from Model VP5 show that the pattern of crustal subduction, detachment, and early exhumation in the high-resolution nested model (Fig. 4) closely resembles that predicted by previous lower-resolution models (Warren et al., 2008a,b). We briefly review this early model evolution before presenting a more detailed discussion of late-stage UHP exhumation and its consequences for upper-crustal structure. The reference model uses $V_p = 5\text{ cm s}^{-1}$; the effect of varying convergence velocity is discussed below with respect to the Tso Morari complex.

Prior to the onset of collision, the subducting oceanic plate undergoes limited rollback, allowing penetration of some sub-lithospheric mantle into the subduction channel (Fig. 4a, f). This produces a weak thermal perturbation and may also lead to incorporation of UHP mantle peridotite into the suture zone (e.g., Dobrzynskay et al., 1996; Robinson et al., 2004); this material may already have had a protracted deep mantle history prior to its emplacement near the base of the lithosphere. Collision, defined as the onset of subduction of the continental margin, begins 10 Ma after the model start; all subsequent model times are reported as Ma-pc (post-collision).

Following collision, the outer margin subducts beneath the accretionary wedge and suture zone formed by earlier-accreted oceanic crust, first reaching UHP conditions at ca. 5 Ma-pc (Fig. 4b). At this stage the subduction channel is narrow (h small) and bulk viscosity is relatively high ($\eta_{eff}$ large), with $E = 0.6$. At about the same time, a low strain nappe ($M^*$, Fig. 4g) starts to form near the top of the channel as shear zones propagate upward through the subducting margin. Between 6 and 8 Ma-pc, a gradually thickening zone of UHP material ($M^*$) accumulates in the lower part of the subduction channel (Fig. 4c). While the volume of buoyant material (h) increases and effective viscosity ($\eta_{eff}$) decreases, thereby increasing $E$ (local $E \sim 4$), the integrated value of $E$ along the channel length is not yet large enough to trigger wholesale exhumation. Near the top of the channel, the $M^*$ nappe becomes decoupled from its substrate (Fig. 4h), forming part of a crustal-scale thrust stack that includes the accretionary wedge. This nappe becomes progressively attenuated between the channel roof and downgoing interior crust (Fig. 4i), but never reaches UHP conditions.

As $h$ increases and $\eta_{eff}$ gradually decreases, sluggish exhumation of $M^*$ begins at ca. 8.5 Ma-pc. The leading edge of the strong continental interior reaches UHP conditions at ca. 9 Ma-pc (Fig. 4d), coinciding with a significant increase in $h$ and therefore exhumation rate ($E \sim 32$). As it exhumes, $M^*$ evolves into a rapidly rising UHP plume ($E \sim 55$) that reaches the lower crust by 10 Ma-pc, underthrusting the tail of the $M^*$ nappe and bowing it upward. A structural dome forms between 10 and 10.5 Ma-pc, cored by the UHP plume and its $M^*$ carapace, and overlain by the deformed suture zone and accretionary wedge (Fig. 4e). By 11 Ma-pc the top of the UHP plume lies within 10 km of the surface. Amplification of the structural dome at 10–11 Ma-pc is accompanied by extension and thinning of the overlying upper crust and coeval foreland-directed thrusting (Fig. 4j). A representative Pht path for exhumed UHP material is shown in Fig. 5f.

The model shows how an upper-crustal accretionary wedge interacts with an underlying subduction channel through which material rises buoyantly on achieving the threshold exhumation number ($E>1$) over the channel length. The rear of the wedge is destabilized by the buoyant plume penetrating its base, leading to extension and normal faulting superimposed on the pre-existing thrust stack, and driving foreland-directed thrusting at the toe of the wedge (Fig. 4j). The exhumed UHP dome takes the form of a nappe stack, cored by the strongly deformed UHP plume, successively overlain by the $M^*$ nappe, highly attenuated remnants of the suture zone, and the extended accretionary wedge. Pressure gaps across major tectonic boundaries reflect juxtaposition of different materials in the upper part of the subduction channel. The dome is bounded on both sides by normal-sense shear zones (Fig. 4j) which thin the carapace and serve to amplify contrasts in geology and metamorphic grade.

Model upper-crustal structures (Fig. 4e,j) closely resemble those reported from many UHP terranes (1–6, Fig. 1), demonstrating that the model produces geologically realistic results. The dominant feature of the model is a structural dome (1), cored by UHP material, directly underlying the suture zone (3). The dome forms in the final stages of exhumation, during and after creation of a nappe stack (4) by juxtaposition of the rising UHP plume with the $M^*$ nappe in the upper part of the subduction channel. As the dome approaches the surface, the nappe stack (4) and overlying upper crust (2) and suture zone (3) become highly attenuated. The rising dome forces extension of the overlying accretionary wedge (6), leading to coeval foreland-directed thrusting (5). Extension and flattening of the $M^*$ nappe and overlying upper crust are driven by the rising UHP plume and are therefore the result, rather than the cause, of UHP exhumation.

5. Comparison with data from the Tso Morari complex

Compatibility of the model with the general structural style of UHP complexes has been demonstrated above. Here we examine a specific natural example by comparing model results with a recent synthesis of data from the Tso Morari complex in the Ladakh Himalaya (Epard and Stock, 2008). This is interpreted as a domal nappe stack, cored by the
UHP Tso Morari nappe (Fig. 5a), derived from rocks of the north Indian continental margin (de Sigoyer et al., 2004; Epard and Steck, 2008). Geological constraints that should be matched by a successful model include the relative times of collision and UHP metamorphism, peak pressure and temperature conditions, time and rate of exhumation to crustal levels, and upper-crustal structural setting.

The time of the India–Asia collision, while debated, is bracketed by deposition of early syn-orogenic deep-water sediments at ca. 55 Ma (Clift et al., 2002; Guillot et al., 2003; DeCelles et al., 2004) and the end of marine sedimentation at ca. 50 Ma (Guillot et al., 2003; Najman et al., 2005; Zhu et al., 2005; Green et al., 2008) (C1, C2, Fig. 5d). The initially high convergence velocity (≥15 cm a⁻¹) decreased significantly sometime between ca. 55 Ma and 48 Ma (Guillot et al., 2000, 2004; Epard and Steck, 2008). The initially overlying Tetraogal nappe was affected by amphibolite facies metamorphism (ca. 550–700 °C, 0.8–1.2 GPa) with top-to-the-south fabrics formed at ca. 48–45 Ma (de Sigoyer et al., 2000, 2004). It lacks evidence for either (U)HP assemblages or the intense deformation associated with the earliest stage of UHP exhumation. The nappe stack now occupies the core of a dome that developed between 47–30 Ma (Epard and Steck, 2008) (Fig. 5a), coeval with SW-directed thrusting in the frontal part of the North Himalayan nappes. The complex is separated from ophiolitic rocks of the Indus Suture Zone by a late normal fault, and dismembered ophiolitic rocks are also present at mid-structural levels within the nappe stack (Epard and Steck, 2008).

Given the high initial convergence velocity during the early stages of the India–Asia collision (Klootwijk et al., 1992; Guillot et al., 2003), we compare the Tso Morari observations with Model V15-5 (the Tso Morari model, Table 1; Fig. A1). This model is exactly the same as the reference Model V1-5 except that V2 = 15 cm a⁻¹ for the first 5.5 Ma of model evolution, representing the initial high convergence rate. Collision begins at 3.3 Ma (0 Ma-pc), crustal thickness in the collision zone increases to 35–40 km by 1.2 Ma-pc, with a shallow foreland basin persisting until the end of the model. UHP metamorphic conditions are first achieved in the subducted margin at 1.5 Ma-pc, and Vp decreases to 5 cm a⁻¹ at 2.2 Ma-pc to represent the post-collision decline in convergence velocity. Exhumation begins at 4.3 Ma-pc, and is well advanced by 5.7 Ma-pc. Beginning at ca. 4.7 Ma-pc, the accretionary
wedge and M′ nappe are deformed into a structural dome above the exhuming UHP plume (Fig. 5b). By 6.7 Ma-pc, UHP material lies in the core of the dome within 10 km of the model surface.

The geometry of the model upper crust at 6.2 Ma-pc corresponds very well to the Tso Morari cross-section (Fig. 5a, b). The UHP dome (M′, Fig. 5c), corresponding to the Tso Morari nappe, is capped by lower-pressure, highly attenuated margin material (M″), corresponding to the Tetraogal nappe, and overlain by low-grade upper-crustal rocks corresponding to the Mata nappe. The dome is bounded by normal-sense shear zones that are coeval with foreland-directed thrusts (Fig. 5c). Within uncertainties, the model is also consistent with available age constraints (Fig. 5d). The onset of collision in the model (0 Ma-pc) has been aligned with the earliest zircon date from the Tso Morari complex (Leech et al., 2007) (U2), which also lies within the range of estimates for the onset of the Himalayan collision (C1). The earliest UHP metamorphism (1.5 Ma-pc) falls within the range of U–Pb zircon (Leech et al., 2005) (U1) and whole-rock–mineral isochron ages (de Sigoyer et al., 2000) (U3), and the initiation of exhumation (4.3 Ma-pc) is consistent with the inferred short duration of UHP metamorphism (O’Brien, 2006). Exhumation of the UHP plume to mid-crustal levels by ca. 6 Ma-pc is compatible with the age of the Tso Morari amphibolite facies overprint (de Sigoyer et al., 2000) (X1, X2). Representative model PT paths (A, B; Fig. 5e) reproduce observed peak PT conditions, duration of UHP metamorphism (< 1 Ma), and most of the exhumation path, although model amphibolite facies temperatures are somewhat too cool. Model exhumation rates (zmax to z ≈ 15 km) are ca. 3 cm a−1 for point A and ca. 1.5 cm a−1 for point B.

The exhumation processes and upper-crustal geometries in Model Vp15-5 (Fig. A1) are very similar to those produced by constant-velocity Model Vp5 (Fig. 4); results from equivalent high-velocity Model Vp15, in which Vp = 15 cm a−1 throughout the model, are presented in Fig. A2. However, the timescales for models Vp5 and Vp15 are too long and too short, respectively, by comparison with Tso Morari data. In addition, although PT paths in all three models reach observed UHP conditions (Fig. 5e, f), both constant-velocity models yield exhumation paths that are too cold. In Model Vp15 this reflects the high convergence rate, which keeps the subduction zone cool; in Model Vp5 this reflects a thinner detached UHP plume, which therefore cools significantly during its ascent. None of the models predicts significant tectonic overpressure (Pconf ~ Pmax). Although decreasing Vp contributes to increasing E (Fig. 2), the models demonstrate that a reduction in subduction velocity is not required to initiate exhumation of UHP terranes. In Model Vp15-5, the velocity decrease was introduced for consistency with inferred changes in Himalayan convergence velocity, and also produces a better match with the observed timing constraints.

6. Discussion and conclusions

The results presented here are significant in two ways. Firstly, they imply that the exhuming UHP plume drives deformation of the overlying near-surface accretionary wedge, producing the characteristic upper-crustal structures of UHP terranes (1–6, Fig. 1), including the Tso Morari complex (Fig. 5). These characteristics are consistent with a single pulse of buoyant exhumation that intrudes and disrupts the accretionary wedge, as demonstrated by Model Vp5 (Fig. 4e). Secondly, the results have implications for the viability of other proposed mechanisms for early collisional exhumation of UHP terranes, particularly in regard to whether they correctly predict observed upper-crustal structures. Both of these aspects are discussed below.

Our high-resolution geodynamical model for the formation and exhumation of UHP rocks during the transition from oceanic subduction to continental collision reproduces key structural and petrological characteristics of early collisional UHP terranes, including a wide range of data from the Tso Morari complex. Key features explained by the present model include: 1) the tectonic setting of UHP complexes near the rear of the accretionary wedge, at high structural level immediately beneath the suture; 2) a single pulse of rapid exhumation within 10 Ma of initial collision; 3) formation of a structural dome above the rising UHP plume, and resulting superposition of normal-sense on thrust-sense structures; and 4) the characteristic pressure gap between UHP rocks and associated nappes, resulting from juxtaposition of the rising plume beneath detached material in the upper subduction channel, and accompanying extension (Fig. 6).

Exhumation is controlled by the characteristic exhumation number, E, which for positively buoyant materials evolves to E > 1 with increasing h and decreasing ηeff in the subduction channel (Fig. 6a). Buoyancy is a necessary but not sufficient condition for rise of UHP material to the upper crust. Near-surface extension and doming are caused by penetration of the buoyant UHP plume (Fig. 6b), which continues to be driven past its point of neutral buoyancy in the crust by the buoyancy force applied by underlying, more deeply subducted crustal material. The substantial buoyancy force required implies that large UHP domes are underlain by relatively thick (≥ 2 km) plumes comprising formerly subducted crustal material. The model explains a broad range of geological observations from early collisional UHP terranes, and significantly advances our understanding of exhumation mechanisms by linking upper-crustal structures to processes operating deep in the subduction channel.

We emphasise that this mechanism can operate in a range of situations where variations in geometry, material properties, and thermal structure lead to differences in the maximum PT conditions and geochronological history recorded by exhumed UHP material. This is because metamorphic P and T, while influencing buoyancy and viscosity respectively, are not the primary controls on exhumation process or rate. Using lower-resolution models, Warren et al. (2008a, b) produced similar exhumation structures for a range of parameter values, with some model PT paths displaying peak conditions near the upper end of the observed natural range (Pmax > 4 GPa, Tmax > 800 °C). In the models, variations in E resulting from initially weaker crustal materials or greater strain weakening may lead to the detachment of crustal rocks at shallower depths in the channel, leading to lower peak pressures. Similarly, the scale of the subduction channel, particularly its depth as

![Fig. 6. Processes involved in formation and exhumation of UHP metamorphic rocks.](image-url)
defined by the thickness of the lithosphere (e.g., normal vs. cratonic), could lead to maximum pressures that vary by up to a factor of two. Maximum temperatures are affected by a number of factors including thermal parameters, material strength, and convergence velocity (e.g., Gerya et al., 2008; Warren et al., 2008a). The natural variability in \( P/T \) paths recorded by UHP complexes that nevertheless display similar upper-crustal structural styles (e.g., Kokchetav, Kaneko et al., 2000; Dabieshan; Fauré et al., 2003) reinforces the interpretation that the burial and exhumation processes investigated here are broadly applicable. In contrast, the same mechanism is unlikely to apply to examples where the timing, duration, and inferred geometry of UHP metamorphism and exhumation differ significantly from the early collisional case considered here (e.g., Western Gneiss Region; Hacker, 2007; Kylander-Clark et al., 2009).

The model results place constraints on the exhumation mechanism by demonstrating that pulse-like buoyant exhumation from deep in the subduction channel correctly reproduces observed upper crustal structures. An obvious question is whether other proposed mechanisms also satisfactorily predict the upper-crustal features of UHP terranes. Mechanisms such as continuous corner flow circulation in a lithospheric-scale wedge or continuous return flow in an over-pressured subduction channel (e.g., Cloos and Shreve, 1988a,b; Platt, 1993; Gerya et al., 2002; Yamato et al., 2008) are considered unsatisfactory for the following reasons. Continuous circulation models are not compatible with pulse-like exhumation of UHP material in discrete events during which the accretionary wedge undergoes extension driven by basal intrusion. All things being equal, continuous circulation in wedges and subduction channels will be accompanied by similarly uniform processes in the upper levels of the accretionary wedge. For example, the wedge should be continuously tectonically resurfaced by the circulation (Platt, 1993), in contrast to observations from early collisional UHP terranes. Similarly, continuous circulation will not reproduce the characteristic pressure gap between the UHP material and overlying nappes, nor will it lead to transient, large-scale, normal-sense faulting in the hanging wall of the exhuming nappe stack. Only pulse-like exhumation can lead to accretionary tectonics characterized by thrusting and stacking of nappes from different levels in the subduction channel, followed by a short-lived extension and doming event driven by emplacement of the exhuming nappe stack into the base of the accretionary wedge.

Other proposed mechanisms, such as exhumation of a strong in-tact slab of the subducted margin (e.g., Chemenda et al., 1995, 2000; Hacker, 2007; Kylander-Clark et al., 2009), or exhumation associated with delamination (e.g., Willner et al., 2002) or transient hot channels (e.g., Gerya et al., 2008), predict exhumation pulses, but their ability to produce acceptable upper-crustal structures has yet to be demonstrated. For example, can this style of exhumation reproduce the characteristic pressure gap between the UHP material and the overlying nappes, or the structural position of the UHP rocks directly beneath the suture? Similarly, the role of slab rollback or breakoff (e.g., Davies and von Blanckenburg, 1995; Kurt and Froitzheim, 2002; Hacker, 2007) in promoting pulses of exhumation cannot be rejected, although present and recent (Warren et al., 2008a,b) modeling results indicate that these processes are not required. Finally, the ubiquitous presence of extensional structures in the upper levels of UHP terranes has led some authors to suggest that crustal extension is a dominant exhumation mechanism (e.g., Andersen and Jamtveit, 1990; Platt, 1993; Johnston et al., 2007). However, in the early collisional setting investigated here, upper-crustal extension is the result, rather than the cause, of UHP exhumation.

We conclude that the formation of structural domes cored by UHP nappes, along with other associated structures (Fig. 1), is an intrinsic part of early collisional UHP rock exhumation. If so, the upper-crustal structures of UHP terranes provide an important additional constraint in the array of observations that must be explained by any successful exhumation model. Acceptable models of UHP metamorphism and exhumation should not only bury and exhum e UHP rocks along appropriate \( P/T \) paths, but should do so in a manner that is consistent with the resulting near-surface geology.

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Appendix A

A.1. Explanation of exhumation number, \( E \)

A.1.1. Interpretation of model behaviour in terms of subduction channel flow and \( E \)

As discussed in the text and previous work (Warren et al., 2008a,b), viscous flow in a subduction channel may be analyzed using lubrication theory (England and Holland, 1979; Cloos, 1982; Cloos and Shreve, 1988a,b; Mancktelow, 1995; Raimbourg et al., 2007). Under the lubrication approximations (Pozrikidis, 2001) the channel flow velocity is:

\[
u(x, y) = -\frac{1}{2\eta} \frac{\partial P_{\text{eff}}}{\partial y} (y - y') + U(1 - y / h) \quad (A1)
\]

where \( \eta \) is the assumed uniform viscosity, \( \partial P_{\text{eff}} / \partial x \) is the effective down-channel pressure gradient (\( x \) measured in the down-dip direction), \( y' \) is the position in the channel measured normal to the base, \( h \) is channel thickness, and \( U \) is the subduction velocity of the underlying lithosphere (Fig. 2). The overlying lithosphere is assumed to be stationary.

Using non-dimensional variables \( u' = u / U, h' = h / h, y' = y / h \) and \( x' = x / h \), Eq. (A1) reduces to:

\[
u' = - \frac{h'^2}{E} \left( y' - y'^2 \right) / 2 + (1 - y'), \quad (A2)
\]

where

\[
E = h^* \frac{\partial P_{\text{eff}}}{\partial x} / \eta_{\text{eff}} U, \quad (A3)
\]

\[
h^* = \left[ 2(\eta_{\text{eff}}) / (\partial P_{\text{eff}} / \partial x) \right]^{1/2}, \quad (A4)
\]

where \( \eta_{\text{eff}} \), the characteristic channel viscosity, and \( \partial P_{\text{eff}} / \partial x \) are scale values used to estimate \( h^* \). This latter parameter is the characteristic channel thickness for \( E = 1 \), the balance point between downward and return flows.

The exhumation number, \( E \) (Warren et al., 2008a,b), equivalent to \( \alpha \) of Raimbourg et al. (2007), expresses the competition between the up-channel Poiseuille flow induced by the pressure gradient and the down-channel Couette flow caused by the viscous drag (Eq. (A2)). In the absence of significant tectonic over- or under-pressure, the effective pressure gradient is a function of the density difference between the material in the channel, \( \rho_c \), and the surrounding lithosphere, \( \rho_m \), such that:

\[
\alpha \frac{\partial P_{\text{eff}}}{\partial x} = (\rho_m - \rho_c) g \sin \gamma \quad (A5)
\]

where \( g \) is the dip of the subduction channel. When \( \rho_m < \rho_c \) the effective pressure increases down the channel, \( \partial P_{\text{eff}} / \partial x \) and \( E \) are positive, and
buoyancy tends to drive upward flow in the channel (Eq. (A3)). The converse always results in overall downward flow and no exhumation.

The behaviour of the simplified channel model can be used to gain insight into the more complex numerical model results by considering $E = E(x')$, such that the material properties and flow regime may vary with position along the channel. Although this approach does not explicitly consider the effect of power law or plastic flows, it does show how different types of flow regime may be established. For steady-state conditions the material flux in the channel, $Q(x')$, is:

$$Q'(x') = -E(x')h(x')/12 + h^2(x')/2 = \text{constant} \quad (A6)$$

with the two terms corresponding to the Poiseuille and Couette components, respectively. Eq. (A6) shows that variations in $E(x')$ can produce differing flow regimes in which the Poiseuille and Couette components vary with depth under conditions of constant flux. Changes in $E(x')$ are accompanied by changes in $h(x')$ such that flow constrictions, stagnant regions, and return flows develop to maintain the constant flux condition ([Raimbourg et al., 2007]). In this simple model $E(x')$ is the only control on the system apart from the boundary conditions and the channel geometry.

Any upward component of flow requires $E>0$, which in turn requires that $\rho_i < \rho_{ma}$, as noted above. Exhumation is also favoured if the downward drag stress, $\eta(x')h^3$, is small, which occurs for combinations of small $U$, low $\eta_{fl}$, and large $h^3$. The larger the positive value of $E(x')$, the greater the exhumation velocity and flux (Eqs. (A2), (A6)). However, as discussed in the text, $E(x')$ should be regarded as a measure of the local exhumation potential; even where the local threshold value is exceeded ($E>1$), efficient exhumation may be impeded by constrictions (low $h$) or high viscosities ($\eta_{fl}$) further up the channel.

**A.1.2. Estimates of $E$ from the numerical models**

We have used plots of velocity, effective viscosity and material properties from the numerical models to estimate local values of the exhumation number in the subduction channel as the models evolve. The list below gives estimates of the critical effective viscosity, $\eta_{fl}$, the value for which $E = 1$ given other parameter values, $h$, $\partial P_{\text{eff}}/\partial x$, and $U$.

For $U = 1$ cm a$^{-1}$, $h = 1000$ m, and $\partial P_{\text{eff}}/\partial x = 2640$ Pa m$^{-1}$ (maximum value for $\rho_i = 2860$ kg m$^{-3}$), $\eta_{fl} = 8.32 \times 10^{19}$ Pa s; when $h = 3160$ m, $\eta_{fl} \sim 1 \times 10^{10}$ Pa s; when $h = 3160$ m and $U = 5$ cm a$^{-1}$, $\eta_{fl} \approx 2 \times 10^{18}$ Pa s; when $h = 6300$ m and $U = 5$ cm a$^{-1}$, $\eta_{fl} \approx 8 \times 10^{14}$ Pa s; and when $h = 10000$ m and $U = 15$ cm a$^{-1}$, $\eta_{fl} \approx 5 \times 10^{20}$ Pa s.

From these estimates it can be seen that $\eta_{fl} < 2 \times 10^{18}$ Pa s is required for the onset of exhumation in Model $V_5$, when detached material in the channel is relatively thin, $h \sim 3$ km (6 Ma-pc). Later (10 Ma-pc), when the detached material in the channel becomes thicker, $h \sim 6$ km, exhumation is possible for $\eta_{fl} < 8 \times 10^{18}$ Pa s. Correspondingly, for $h = 10$ km (10 Ma-pc), exhumation can occur when $\eta_{fl} < 1.66 \times 10^{21}$ Pa s. It can also be seen that moderate changes in the bulk density of the upper/midcontinental crust during metamorphism have only a minor effect on $\partial P_{\text{eff}}/\partial x$ and correspondingly little effect on $E$. The increasing value of $E$ is primarily a consequence of increasing $h$ and decreasing $\eta_{fl}$. Values of $E$ estimated from Model $V_5$ are shown on Fig. 4 and the corresponding ones for Model $V_2$ and Model $V_5$ are shown in Figs. A1 and A2.

**A.2. Methods**

**A.2.1. Explanation of nested version of the numerical model**

Sopale Nested is a version of the plane strain ALE finite element software Sopale ([Fullsack, 1995]) that solves thermal-mechanical creeping flows in a large-scale (LS) domain in which there is embedded a second small-scale (SS) sub-domain. Both the LS and SS models are defined in the SS domain. The purpose is to achieve a higher-resolution solution by solving the problem sequentially, first for the LS domain and then for the high-resolution SS domain using boundary conditions derived from the LS solution. Under normal circumstances the finite element grid for the SS domain is defined by increasing the resolution of the LS grid by multiples $m$ and $n$ in the horizontal and vertical directions. This means that the LS grid in the SS domain is the SS grid reduced by removing all but multiples of every $m$th and $n$th column and row.

The two domains share a single cloud of Lagrangian tracking or tracer particles. The initial positions of these particles are nearly coincident with the nodes of the LS finite element grid and at the same density within model regions outside of the LS domain. Additional particles are added such that $a \times b$ particles are proportionately positioned within each SS element, and the same distribution is extended throughout the SS domain. The density of particles, i.e., $ma$ and $nb$, is chosen so that the model geometry and properties can be advected through both the LS and SS domains without loss of fidelity.

The initial conditions are defined for the entire model and are then transferred to the LS and SS domains. For each model time step the problem is first solved iteratively on the LS domain using LS boundary conditions. The velocity and temperature from the LS solution at nodes that correspond to the boundary of the SS domain are linearly interpolated onto the boundary nodes of the SS domain, and the problem is solved iteratively for the SS domain using these boundary conditions. Finally, the particle properties are updated and their positions advected using the LS and SS solutions for their respective domains.

Coupling between the LS and SS domains is achieved by the replacement of the LS solution by the SS solution in the relevant part of the LS domain. The solutions do not drift apart because model properties are defined and transported by a single Lagrangian cloud of particles. Particles used to track the evolving properties of the model, for example those used to record pressure–temperature–time (P–T) paths, are drawn from the Lagrangian cloud. They will therefore record the LS and SS solutions when in these respective domains. Where part of the SS domain is a free surface, the reduced SS free surface solution replaces the equivalent LS solution for that surface.

In the current models, the position of the SS domain does not move with time. However, there is no restriction on the position of the SS domain for each time step so that various schemes can be employed to move the SS domain in any desired manner.

This nested, dual, finite element modeling technique is accurate and robust when the LS solution provides accurate boundary conditions for the SS domain. This means that the velocity and temperature fields must therefore be determined with sufficient accuracy by the LS solution in the area outside the SS domain. Computationally, the problem is solved using separate versions of Sopale on two processors. The exchange of information required to implement the steps outlined above is achieved by message passing. In principle, multiple SS domains can be defined within an LS domain and SS domains can be nested within each other like Russian dolls.

The nested model results represent a significant increase in resolution ($2 \times 2$ km finite elements within SS) by comparison with earlier results ($10 \times 2$ km finite elements within LS). Nevertheless, model features such as the subduction channel, bounding shear zones, and exhuming UHP material may still be wider than their natural counterparts.

**A.2.2. Equations solved, and boundary and initial conditions**

The equations for incompressible creeping (Stokes) flows (Eqs. (A7) and (A8)) and energy balance Eq. (A9) are solved in the large-scale (LS) and small-scale (SS) model domains:

$$\frac{\partial \sigma_{ij}}{\partial x_i} + \rho g = 0 \quad i, j = 1, 2. \quad (A7)$$

$$\frac{\partial v_i}{\partial x_i} = 0 \quad i = 1, 2. \quad (A8)$$

$$pc_{ij} \left( \frac{\partial T}{\partial t} + v_i \frac{\partial T}{\partial x_i} \right) = \frac{\partial}{\partial x_i} K \frac{\partial T}{\partial x_i} + A + A_{gh} + v_i c_{gh} T \rho \quad i = 1, 2. \quad (A9)$$

where $c_{ij}$ is the deviatoric stress tensor, $x_i$ are the spatial coordinates, $P$ pressure, $\rho$ density, $g$ gravitational acceleration, $v_i$ a component of
velocity, \( c_s \), specific heat, \( T \) temperature, \( t \) time, \( K \) thermal conductivity, \( A \) radioactive heat production per unit volume, \( A_{\omega} \) shear heating, and \( \alpha \) volumetric expansivity. Most of the parameters (e.g., \( \rho, K, A \)) vary with the type of material (Table 1). The last term in the heat balance equation is the temperature correction for adiabatic heating when material moves vertically at velocity \( v_s \). During phase transitions incompressibility (A8) is replaced by mass conservation (A2.5).

The initial steady-state temperature field is calculated at the model scale, with 0 °C surface temperature, insulated side boundaries, radioactive heating, and a basal heat flux (Fig. 3, Table 1). For the continental interior this gives Moho temperature (570 °C), basal lithosphere temperature (1336 °C), and corresponding surface heat flow (55 mW m\(^{-2}\)), typical of continental lithosphere (e.g., Rudnick and Fountain, 1995). The sublithospheric thermal conductivity (Table 1) has a high value to maintain a heat flux equivalent to the average that would be transported convectively. This also ensures that the models maintain a near-adiabatic sublithospheric thermal gradient as would be expected in a convecting system.

Pro-continental lithosphere converges on the stationary retro-lithosphere at rates that vary among the models from 5 to 15 cm\(^{-1}\) (Fig. 3). The upper boundary of the model is a stress-free surface. The sides and base of the sub-lithospheric mantle domain have no-slip and free-slip boundary conditions respectively. Entry of lithosphere into the model domain is volume-balanced by a uniform outward low-velocity flow through both side boundaries below the lithosphere. This leakage flow is further modulated to keep the model in constant isostatic equilibrium by maintaining an average pressure at the base of the model. This ensures that the model surface remains vertically balanced with the larger-scale upper mantle outside the model domain.

A2.3. Material properties

Our general approach is to keep model parameterization as simple as possible in order to facilitate interpretation of model results. This obviously introduces uncertainties into the models, but we consider these to be of similar order to the uncertainties involved in our understanding of natural systems. Major simplifications include the choice of rheological flow laws and their relative scaling, and constant, non-temperature-dependent thermal conductivity. Model sensitivity to the choice of margin width, margin heat production, collision velocity, and strain level for the onset of strain weakening was investigated previously (Warren et al., 2008a,b). Here we use wet quartzite (WQ; melt-absent Black Hills quartzite; Gleason and Tullis, 1995), dry Maryland diabase (DMD; Mackwell et al., 1998), and wet olivine (WO; Karato and Wu, 1993). The values of \( B^* \) are scaled linearly by a scaling factor \( f \) (Eq. A12), which is fixed for most materials but variable for the continental margin, to represent lithologies that are stronger or weaker than the base set, wet vs. dry conditions, or moderate changes in composition (Beaumont et al., 2006; Warren et al., 2008a,b). This scaling helps to minimize the number of sources of error while allowing some variation in flow properties, simplifies the interpretation of model results, and acknowledges uncertainties in the composition and rheological properties of the Earth. The choice of reference flow laws (Table 1) does not imply that the corresponding earth material has the composition of the reference material.

Margin upper/mid crust (Fig. 3b) was chosen to be WQ, \( f = 0.4 \), to approximate a quartz-controlled rheology somewhat weaker than WQ (Gleason and Tullis, 1995). This can be interpreted either as crust with a significant sedimentary or other weak component, or as a measure of the range of uncertainty in the WQ flow law (Gleason and Tullis, 1995) which may predict relatively high viscosities (Hirth et al., 2001; Burov, 2003; Kenis et al., 2005). Continental interior upper/mid crust is WQ with \( f = 5 \) (Fig. 3b), to represent more refractory continental crust. Similarly, the rheology of the model lower crust (DMD scaled by \( f = 0.1 \)) corresponds closely to flow laws for intermediate granite (Mackwell et al., 1998).

A2.4. Strain softening and weakening of crustal materials

Crustal materials weaken in both the frictional-plastic and ductile regimes (Warren et al., 2008a,b; Huismans and Beaumont, 2003; Sobolev and Babeyko, 2005), termed “softening” and “weakening”, respectively. Frictional-plastic materials strain-soften through a linear decrease in the effective internal angle of friction, \( \phi_{\text{eff}} \), with accumulated plastic strain: \( \phi_{\text{eff}} = \phi_{\text{eff}}(l_2^{1/2}) \). \( l_2^{1/2} = \epsilon \), is the square root of the second invariant of deviatoric strain, with \( \epsilon \) (“strain”) used for simplicity. This approach approximates deformation-induced mechanical or pore-fluid pressure softening of faults and brittle shear zones. In all model materials apart from the weak seed, \( \phi_{\text{eff}} \) is reduced from 15° to 2° over 0.5 \( \leq \epsilon \leq 1.5 \) (Huismans and Beaumont, 2003). The weak zone (Fig. 3a) is modeled as an initially weak region, \( \phi_{\text{eff}} = 5^\circ \), inherited from earlier deformation.

In natural rocks, localized regions of deformation (faults and shear zones) are ubiquitous on microscopic to macroscopic scales. Exhumed (U)HP terranes have been described as pervasively deformed (Terry and Robinson, 2004), or as low-strain regions separated by high-strain shear zones (Jolivet et al., 2005). How, when, and where these high-strain zones initiate and propagate are poorly constrained, but experiments suggest that they are significantly weaker than the host rock and remain weak once formed (Holyoke and Tullis, 2006). The amount of weakening is, among other factors, variably dependent on mineralogy, temperature, and fluid composition and distribution.

In the models, viscous strain weakening, \( W_p \), proceeds through a linear decrease in effective viscosity by the specified factor over a specified \( \epsilon \) range. Model sensitivity to \( W_p \) and \( \epsilon \) range has been investigated previously (Warren et al., 2008a,b). Here we use \( W_p = 10, 5 \leq \epsilon \leq 10 \), as a proxy for weakening owing to a combination of effects including grain size reduction and reaction weakening (e.g., from hydration).

Low values of strain weakening, \( 2 \leq W_p \leq 5 \), can be related to those observed in laboratory experiments (Bystricky et al., 2000; Heidelbach
et al., 2001; Barnhoorn et al., 2004) and to grain-size reduction during mylonite development. The range of strain, \(5 \leq \varepsilon \leq 10\), over which viscous weakening operates in the models, is the same for all materials that strain-weaken and does not vary among models. We use \(W_s = 10\) to represent the effect of strain weakening, possibly compounded by reaction weakening, in particular the effects of hydration (Warren et al., 2008a). By linking the two mechanisms we imply that fluids are available, but that their access is facilitated by deformation and is therefore linked to increasing levels of strain.

A.2.5. Density, volume, and mass conservation during phase transitions

Lagrangian particles, embedded in the LS continental margin and interior, are tracked to monitor \(PT_t\) evolution. Model crustal materials increase and decrease in density and volume at \(PT\) conditions corresponding to metamorphic phase changes (Table 1; Warren et al., 2008a). The densities of lower continental and oceanic crust change reversibly across the eclogite field boundary, from 2950 to 3100 kg m\(^{-3}\) and 2900 to 3300 kg m\(^{-3}\) respectively (Table 1). The higher density of oceanic-crust eclogite reflects its mafic composition, whereas the intermediate lower continental crust is assumed to have a lower eclogite–facies density. The density of upper- and mid-crustal material, assumed to consist of 10% mafic and 90% felsic material by volume, changes from 2800 to 2850 to 2900 kg m\(^{-3}\) through the eclogite andcoesite–eclogite field boundaries respectively. The calculated densities assume that all mafic material but only 20% of the felsic material transforms to denser phases, with all quartz transforming to coesite across the quartz–coesite phase boundary. Density changes reverse during exhumation.

For materials that change density during a phase change, the incompressibility equation (A8) (A2.2) is modified to that of mass conservation: \(\frac{\partial \rho}{\partial t} = -\frac{\partial (\rho v_i)}{\partial x_i}\). This accounts for the associated volume change and its effect on the buoyancy and velocity field. This volume change is calculated numerically by applying additional normal, compressive/dilatational forces to finite elements at the time they are subject to phase-related density changes. The value of the excess pressure is \(\Delta P = \Delta \rho/\beta_v\), where \(\beta_v\) is the viscous bulk modulus of the material, and \(\Delta \rho/\rho\) is the fractional change in density corresponding to the phase change. The excess pressure compresses material locally and only during the model time steps when the phase changes occur, thereby ensuring mass conservation. The fractional volume change accompanying a phase change is small in these models and its effect on the velocity field is minor because it only applies at the time of the phase change. However, failure to ensure mass conservation has a long-term effect on the model because the buoyancy forces will be over- or under-estimated by the fractional error in the material volume.

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**Fig. A1.** Selected frames from Model Vp15-5, in which convergence velocity decreases from 15 cm a\(^{-1}\) to 5 cm a\(^{-1}\) at 2.2 Ma-pc. This is the preferred model for comparison with the Tso Morari complex, as discussed in text. Onset of UHP metamorphism at 1.5 Ma-pc is identical to that shown for Model Vp15 (Fig. A2a) and is not shown here. Panels a–c: distribution of model materials at key stages in model evolution. \(E = \) estimates of exhumation number in outlined region; see A1.2 for details. Panels d–f: strain at same times. A representative \(PT_t\) path from this model is shown in Fig. 5f (main text). Animations in Supplement (Vp15-5_materials.mov, Vp15-5_strain.mov).
Appendix B. Supplementary data

Supplementary data associated with this article can be found, in the online version, at doi:10.1016/j.epsl.2009.08.001.

References


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Fig. A2. Selected frames from Model Vp15, with constant convergence velocity, $V_p = 15$ cm a$^{-1}$. Panels a–c: distribution of model materials at key stages in model evolution, as discussed in text for reference Model Vp5. Onset of UHP metamorphism at 1.5 Ma-pc is identical to that for Model Vp15-5 (not shown in Fig. A1). E = estimates of exhumation number in outlined region; see section A1.2 for details. A representative PTt path from this model is shown in text Fig. 5f. Panels d–f: Strain at same times. Animations in Supplement (Vp15_materials.mov, Vp15_strain.mov).


