Crustal Flow Modes in Large Hot Orogens

CHRISTOPHER BEAUMONT[1], MAI NGUYEN[1, 2], REBECCA JAMIESON[2] & SUSAN ELLIS[3]

[1] *Oceanography Department, Dalhousie University, Halifax, Nova Scotia, Canada B3H 4J1,* chris.beaumont@dal.ca

[2] *Department of Earth Sciences, Dalhousie University, Halifax, Nova Scotia, Canada B3H 3J5*

[3] *Institute for Geological and Nuclear Sciences, Lower Hutt, New Zealand*

Abstract: Crustal scale channel flow numerical models support recent interpretations of Himalayan-Tibetan tectonics proposing that gravitationally driven channel flows of low-viscosity, melt weakened, middle crust can explain both outward growth of the Tibetan plateau and ductile extrusion of the Greater Himalayan Sequence. We broaden the numerical model investigation to explore three flow modes: Homogeneous Channel Flow (involving laterally homogeneous crust); Heterogeneous Channel Flow (involving laterally heterogeneous lower crust that is expelled and incorporated into the mid-crustal channel flow); and Hot Fold Nappes style of flow (in which mid-/lower crust is forcibly expelled outward over a lower crustal indentor to create fold nappes that are inserted into the mid-crust). The three flow modes are members of a continuum in which the Homogeneous mode is driven by gravitational forces but requires very weak channel material. The Hot Fold Nappe mode is driven tectonically by, for example, collision with a strong crustal indentor and can occur in crust that is subcritical for Homogeneous flows. The Heterogeneous mode combines tectonic and gravitationally-driven flows. Preliminary results also demonstrate the existence and behaviour of mid-crustal channels during advancing and retreating dynamical mantle lithosphere subduction. An orogen Temperature-Magnitude (T-M) diagram is proposed and the positions of orogens in T-M space that may exhibit the flow modes are described, together with the characteristic positions of a range of other orogen types.

Abbreviated title: Crustal Flow Modes

During the last decade we have developed and used a range of finite element numerical models to gain insight into collisional orogenesis. These types of models include 2D doubly (bi-)vergent (Willett *et al*. 1993; Beaumont *et al*. 1994; Beaumont & Quinlan 1994), 3D doubly (bi-)vergent (Braun & Beaumont 1995), vise (Ellis *et al*. 1998), accretionary wedge (Beaumont *et al*. 1999), Pyrenean (Beaumont *et al*. 2000), and Alpine styles (Beaumont *et al*. 1996a; Ellis *et al*. 1999; Pfiffner *et al*. 2000). Both mechanical and thermo-mechanically coupled (Jamieson *et al*. 1998) techniques have been applied to small-cold and large-hot orogens (Jamieson *et al*. 2002). The applications to small orogens include the Pyrenees, Alps, Southern Alps of New Zealand (Beaumont *et al*. 1992; Beaumont *et al*. 1996b; Waschbusch *et al*. 1998), and to examples studied by the Canadian Lithoprobe program (Ellis & Beaumont 1999). Applications to large hot orogens include the Himalayas and Tibet (Willett & Beaumont 1994; Beaumont *et al*. 2004; Jamieson *et al*. 2004b).

Each type of orogen occupies a characteristic position in an orogenic Temperature-Magnitude (T-M) diagram (Fig.1). This concept is inspired by the astrophysical Hertzsprung-Russell (H-R) star diagram, in which luminosity (or absolute visual magnitude) is plotted against spectral type (or surface temperature) for star populations (Fig.1a, top right) (Hertzsprung 1905; Russell 1914). The H-R diagram concisely describes stellar conditions and provides insight into the range of stellar evolution. The T-M diagram (Fig.1) is intended to express relationships between the magnitude of the orogen, measured by excess crustal or lithospheric thickness of the orogen relative to that of undeformed standard continental lithosphere, and the excess heat content or temperature of the orogen relative to the same undeformed lithosphere with standard heat production. The T-M diagram provides a first-order classification of orogen types (e.g. Dwarfs, Giants, Fig 1a) and offers insight into the underlying tectonic processes. In addition, the evolution of orogens can be represented by evolutionary paths in the T-M diagram. For example, an accretionary wedge may evolve into a cordilleran orogen and thence to a large continent-continent collisional orogen. An 'orogenic main sequence' (MS, Fig.1) extends from bottom left toward the top right. Orogens on the main sequence have excess conductive steady-state temperatures complementing their excess crustal thicknesses. The main sequence is nonlinear for two reasons. Curvature of the lower part occurs because conductive steady-state temperatures contain a quadratic or higher order term owing to radioactive heat production. Steady state average temperatures in thickened crust/lithosphere increase disproportionately faster than the corresponding thickening of the orogen and, therefore, the MS plots below the $M=\alpha T$ line, where α is a measure of the ratio between magnitude and temperature in the standard state. The convex-up curvature for large orogens expresses the trend to a gravitational limit on the maximum thickness of hot, weak crust and lithosphere. Here the T-M diagram is presented in a one-dimensional form, but even this conceptual view can take account of some three-dimensional aspects. For example, orogens that grow during orthogonal collision accumulate mass more rapidly than equivalent orogens where motion is primarily transcurrent, and the former will evolve more rapidly in T-M space. Transcurrent orogens may never evolve out of the small-cold part of T-M space.

The primary division in a T-M diagram is between small-cold and large-hot orogens (Fig.1b). Small-cold orogens such as accretionary wedges, Southern Alps of New Zealand, Pyrenees, and Alps plot in the lower left part of diagram (Fig.1b); they lack the levels of crustal thickening and associated thermal relaxation necessary to achieve high temperatures. This may be because they are young (Proto-Main Sequence, Fig.1a), strongly denuded (Denudation Dwarfs, Fig.1a), dominantly transcurrent, or have low levels of radioactive heating (Accretionary Coldies, Fig.1a). It follows that their minimum crustal viscosities are too high for large-scale fluid-like flows in the middle and/or lower crust.

At the other end of the orogenic main sequence, the large-hot orogens (Giants and Super Giants, Fig.1a) are both massive and hot, leading to weak viscous regions in the crust that may contain *in situ* partial melts and that may undergo gravitationally-driven channel flows (Bird 1991; Westaway 1995; Royden 1996; Royden *et al*. 1997; Beaumont *et al*. 2001; Shen *et al*. 2001). Such flows of 'melt weakened' crust can explain both eastward growth of the Tibetan plateau, as the channel tunnels outward (Clark & Royden 2000, and references therein), and ductile extrusion of the Greater Himalayan Sequence (Grujic *et al*. 1996, 2002; Beaumont *et al*. 2001, 2004; Jamieson *et al*. 2004b). We regard gravitationally driven channel flow as an end member requiring a combination of sufficiently low viscosities, thick channels, and large differences in mean elevation between the orogen and its foreland to allow the available differential pressures to drive efficient flow (Bird 1991; Clark & Royden 2000). If gravitationally-driven crustal flow is an end-member that exists only in Giant and Super Giant orogens like Tibet (Fig. 1b & a), do other flow modes occur when conditions are subcritical for gravitational forcing? If so, what drives these flows? How do these more general flow regimes relate to the evolutionary paths outlined in Figure 1? We address these questions in this paper.

The mechanics of and types of models used to investigate small-cold orogens (Fig.1c) have been described in the earlier papers cited above. Here we expand on the types of flow that can occur within large-hot orogens. We use numerical models to investigate three modes of crustal flow. Mode 1 is Homogeneous Channel Flow; Mode 2 is Heterogeneous Channel Flow, which incorporates lower crustal blocks within the channel; Mode 3, Hot Fold Nappes, measures the response of the model orogen to the insertion of progressively stronger blocks of lower crust. We interpret these as members of a continuum of gravitationally and tectonically driven flow modes, and relate the results to the corresponding deformation predicted for orogenic crust in the large-hot orogen region of the T-M diagram. We also provide preliminary results from upper mantle-scale models that address the fate of lithospheric mantle during continent-continent collision, and show that channel flows also exist within this model style.

Numerical calculation of crustal- and upper-mantle-scale flows

The numerical modeling methodology is outlined in this section. An explanation for the choice of the model parameter values and the sensitivity of the results to this choice is included in the Appendix. There are two types of models. Crustal-scale (CS) models were described in Beaumont *et al.* (2004) but an explanation is included here for completeness. Upper-mantle-scale (UMS) models are discussed later but their primary properties are described here. In both CS and UMS we model the development of large-hot orogens using a two-dimensional (2D) finite element code that assumes plane-strain conditions in a vertical cross-section through the orogen. The codes compute thermal and mechanical evolution subject to velocity boundary conditions applied at the sides and base of the CS model region, and applied at the sides of the UMS model region. Thermal-mechanical coupling occurs through the thermal activation of viscous power-law creep in the model materials and through the redistribution of radioactive crust by material flow.

The CS model properties are similar to those described by Beaumont *et al*. (2004) and Jamieson *et al*. (2004b). This model, which is 2000 km wide, has two regions - the crust (Fig. 2a & b), in which the velocity and deformation are calculated dynamically; and the mantle, where the velocity is prescribed kinematically (Fig. 2b). In the 2000 x 600 km UMS model, the velocity and deformation for the whole model domain are calculated dynamically subject to far-field lateral velocity boundary conditions. The associated temperature field is calculated for the whole model domain. Model parameters and values for CS and USM models are given in Tables 1 and 2 respectively.

CS model velocity boundary conditions and reference frames

In the CS models described in this paper, both the pro- and retro-mantle lithospheres (Fig. 2) converge at a uniform velocity, $V_P = -V_R$, and detach and subduct beneath the stationary S-point (Willett *et al*. 1993). This frame of reference and associated symmetric convergence were chosen to give results that are least dependent on the motion of the lithospheric plates with respect to the sublithospheric mantle. The subducted mantle lithosphere descends into the mantle at constant dip with constant kinematically specified velocity (Fig. 2). The CS models can be interpreted in other reference frames (Beaumont *et al*. 2004; Fig. 9) by adding or subtracting a fixed velocity to all of the boundary velocities and the velocity of the S-point. For example, Jamieson *et al.* (this volume) investigate the case where the pro-mantle lithosphere converges at $2V_P$, the S-point advances at $V_S = V_P$, and the retro-mantle lithosphere is stationary, $V_R = 0$. This reference frame is considered most appropriate for Himalayan-Tibetan models (Beaumont *et al*. 2001, 2004; Jamieson *et al.* 2004b). The change in reference frame does not change the model results, only the way in which they are viewed.

UMS model velocity boundary conditions and reference frames

The UMS model (Fig.16) is designed to correspond approximately to the collision of India with Asia. The boundary condition has pro-lithosphere, equivalent to India, converging from the left boundary at a uniform velocity of $V_P = 5$ cm/y against a stationary retro-lithosphere, $V_R = 0$ cm/y, at the right boundary corresponding to Asia. In contrast to the CS models, V_S is not specified but is determined by the dynamical evolution of the model. In the sublithospheric mantle region, the sides and base have free slip boundary conditions. A small uniform symmetric outward leakage flux of material is specified through the side boundaries to balance the flux of pro-lithosphere into the model; there is no material flux through the base of the model.

CS and UMS mechanical models

The mechanical models used to calculate the CS and UMS velocity fields and deformation (Fullsack 1995) use an Arbitrary Lagrangian Eulerian (ALE) methodology in which flows with free upper surfaces and large deformation are calculated on an Eulerian finite element grid that stretches in the vertical direction to conform to the material domain. A Lagrangian grid, which is advected with the model velocity field, is used to update the mechanical and thermal material property distributions on the Eulerian grid as their position changes. Flow is driven by the basal and lateral velocity boundary conditions described above.

The finite-element model uses a viscous-plastic rheology. The plastic (frictional or brittle) deformation is modelled with a pressure-dependent Drucker-Prager yield criterion. Yielding occurs when

$$(J_2^{\prime})^{1/2} = P \sin\phi_{eff} + C \cos\phi_{eff} \qquad (1)$$

where $J_2^{\prime}$ is the second invariant of the deviatoric stress, P the dynamical pressure (mean stress), C the cohesion, and the internal angle of friction, ϕ_{eff}, is defined to include the effects of pore fluid pressures through the relation

$$P \sin\phi_{eff} = (P - P_f) \sin\phi \,. \qquad (2)$$

For dry frictional sliding conditions (approximating Byerlee's law), $\phi = 30°$ when the pore fluid pressure, $P_f = 0$. For hydrostatic fluid pressures and typical crustal densities ϕ_{eff} is approximately 15°, and for overpressured pore fluid conditions we use $\phi_{eff} = 5°$ (see Appendix) .

The incompressible plastic flow becomes equivalent to a viscous material (Fullsack 1995; Willett 1999) such that $\eta_{eff}^{P} = (J_2^{\prime})^{1/2} / 2(\dot{I}_2^{\prime})^{1/2}$, where $(\dot{I}_2^{\prime})^{1/2}$ is the second invariant of the deviatoric strain rate. Setting the viscosity to η_{eff}^{P} in regions that are on frictional-plastic yield satisfies the yield condition and allows the velocity field to be determined from the finite element solution for viscous creeping flows. The overall non-linear solution is determined iteratively using $\eta = \eta_{eff}^{P}$ for regions of plastic flow, and $\eta = \eta_{eff}^{V}$, as defined below for regions of viscous flow.

The flow is viscous when the flow stress is less than the plastic yield stress for the local ambient conditions. Under these circumstances the power law creep effective viscosity is

$$\eta_{eff}^{V} = B^{*} \cdot (\dot{I}_2^{/})^{(1-n)/2n} \cdot \exp[Q / nRT_K]. \tag{3}$$

The values of *B**, *n*, and *Q* (Table 1) are based on laboratory experiments with *A* values converted to *B** assuming cylindrical creep tests. The rheology of the upper and middle crust is based on the 'Wet Black Hills Quartzite' *(WQ)* flow law (Gleason & Tullis 1995). In the CS model experiments (Table 1) we use the flow law with *B** = *B*(WQ)* in the uppermost crust (initially 0-10 km). In the mid-crust (initially10-25 km) *B** is scaled by a factor of 5 (*B*=B*(WQx5)*), as explained in the Appendix. In the UMS models the upper and middle crust have *B** = *B*(WQx5)* (Table 2). The rheology of the lower crust (initially 25-35 km) is based on the 'Dry Maryland Diabase' *(DMD)* flow law (Mackwell *et al.* 1998) (Table 1), which is also scaled to achieve a range of effective lower-crustal strengths (Appendix).

The reference CS rheological structure in model LHO-1 represents a laterally uniform three-layer crust. The upper layer (*B*(WQ)*) has weak frictional-plastic properties, ϕ_{eff} =5°. The middle layer (*B*(WQx5)*) has standard hydrostatic frictional-plastic properties, ϕ_{eff} =15°. This is underlain by lower crust with ϕ_{eff} =15° and *B** = *B*(DMD/5)*. This layering is designed to approximate the continental margin crust commonly involved in collisional orogenesis, with a refractory, intermediate granulite lower crust overlain by middle crust comprising fertile quartzo-feldspathic low-grade metasedimentary and granitic rocks and an upper crust dominated by quartz-rich sedimentary rocks with high pore fluid pressures.

In UMS models, the crustal rheology is similar to the CS models (Tables 1 and 2) except that there is no separate weak upper crustal layer. The initial thicknesses of the layers are also slightly different because the Eulerian finite element resolution is lower, comprising 17 as opposed to 40 crustal elements.

The rheology of the mantle in the UMS models is based on the Wet Aheim Dunite (olivine) flow law (*WO*) (Chopra & Paterson 1984) which is similar to that from Karato & Wu (1993) for wet olivine. This flow law is used for the sublithospheric mantle which is considered to be water-saturated ('wet'). The lithospheric mantle in the UMS experiments is assumed to be more refractory and water-poor. The value of B^* is therefore scaled to $B^*(WOx10)$ to represent mantle lithosphere that is stronger owing to lower water fugacity (Appendix). This scaled flow law predicts effective viscosities that are intermediate between the 'wet' and 'dry' olivine-controlled flow laws of Chopra & Paterson (1984) and corresponding results in Karato & Wu (1993). The effect of the activation volume is not included in the calculation of the power-law creep flow laws (Appendix).

There is no strain dependence of the material properties in the CS models described here or in our other papers on large hot orogens (Beaumont *et al*. 2001, 2004; Jamieson *et al.* 2004b, this volume). However, strain-softening is included in the UMS models, in the same parametric manner as described by Huismans & Beaumont (2003), by reducing the value of ϕ_{eff} linearly from 15° to 2° as the second invariant of the deviatoric strain increases from 0.5 to 1.5 (Table 2, Appendix). Strain-softening occurs in all plastic materials but there is no strain-softening during viscous flow.

Melt weakening

The most important additional property in both CS and UMS models is an extra increment of viscous weakening in the upper and middle crustal materials (those based on the *WQ* flow law) such that the effective viscosity decreases linearly with temperature from the dynamically determined power law creep value at $T = 700$°C to 10^{19} Pa.s at $T \geq 750$°C (Fig. 2). This weakening approximates the reduction in bulk viscosity caused by a small amount of *in situ* partial melt, estimated to be ca. 7% at the melt connectivity transition (Rosenberg & Handy 2005). This weakening does not correspond to, and is not designed to represent, the additional decrease in effective viscosity that occurs at much larger partial melt fractions at the solid to liquid transition. The models therefore cannot be interpreted in terms of magma accumulation, transport, or emplacement. The 'melt weakening' used in the present models amounts to approximately a factor of 10 decrease in effective viscosity, probably a conservative estimate for melt weakening by a small percentage of *in situ* melt. The lower crust in the models does not melt weaken because it is interpreted to be refractory intermediate granulite not prone to dehydration melting at the temperatures achieved in the models.

Model materials can therefore deform according to two mechanisms; plastic or viscous flow, and in the latter case the viscosity may be further reduced by melt weakening. In all instances, the material deforms according to the mechanism that produces the lowest level of the second invariant of the deviatoric stress for the prevailing conditions; that is, the weakest of the available flow regimes is chosen.

Density structure and isostatic compensation

In CS models the crust has a uniform density (Table 1); no account is taken of density changes owing to variations in thermal expansion, melting, or metamorphism. This approach is adopted so that buoyancy forces act equally on all materials and none of the flow results from differential buoyancy forces caused by density variations. The changing crustal thickness is isostatically compensated by elastic flexure of a beam embedded in the model at the base of the crust. The flexural rigidity, $D = 10^{22}$ Pa.s, is sufficiently low that broad regions of uniform thickness crust beneath plateau regions in the model are effectively locally compensated. Only at the transition from plateau to undeformed crust is the effect of flexure apparent. Model topography depends on the choice of crustal and mantle densities (e.g., Fig. 4 of Beaumont *et al.* 2004).

The UMS models are more dynamical than their CS equivalents and, consequently, they are more sensitive to their density structure. The upper and middle crust has a uniform density, and the lower crust has a higher density which increases over the P-T range corresponding to the granulite-eclogite transition (Table 2). With the exception of the sublithospheric mantle all materials have a uniform volume coefficient of thermal expansion (Table 2). The sublithospheric mantle has a constant density and is, like the other materials, incompressible. The UMS models are ‘isostatically’ compensated at the scale of the model by their internal density structure and the associated gravitationally-driven component of the flow.

Thermal model

The thermal evolution is calculated by solving the heat balance equation,

$$\rho C_p \partial T / \partial t + \underline{v} \cdot \nabla T = K \nabla^2 T + A \qquad (4)$$

on an Eulerian finite element mesh, where ρ is density, C_p is specific heat, T is temperature, t is time, $\underline{v}$ is the advection velocity of the material, K is thermal conductivity, and A is radioactive heat production per unit volume. In CS models the Eulerian finite element mesh is the same as that for the mechanical model in the crust and continues into the underlying mantle as shown in Fig. 2b. The advection velocities are calculated dynamically in the crust and are prescribed kinematically in the mantle. In UMS models the heat balance equation is solved for the whole model domain using dynamically calculated velocities.

In both CS and UMS models the values of K, ρ (thermal density), and C_p are uniform throughout the model lithosphere, resulting in uniform thermal diffusivity, κ (Tables 1, 2). The upper crust (0-20 km) has a uniform radioactive heat production, $A_1 = 2.0$ μW/m^3, and the lower crust (20-35 km CS, 20-34 km UMS) has lower heat production, $A_2 = 0.75$ μW/m^3 (Jamieson *et al*. 2002) (Tables 1, 2, Appendix).

For each model run, the initial steady-state temperature field is calculated at the scale of the model, with a surface temperature of 0°C and no heat flux through horizontal side boundaries. The basal heat flux, $q_m = 20$ mW/m^2, is applied at the base of UMS models, and at the lithosphere-asthenosphere boundary, defined to coincide with the 1350°C isotherm, in CS models. For these conditions and thermal conductivity $K = 2.00$ W/m°C, the initial surface heat flux $q_s = 71.25$ mW/m^2, and the Moho temperature in CS models is 704°C. These values are slightly lower in UMS models (Table 2) because the crust is 34 km rather than 35 km thick. The effect of a precursor phase of oceanic subduction, included in some of our models (e.g. Vanderhaeghe *et al*. 2003), is not included because it has little effect on the evolving crustal temperatures and peak metamorphic conditions at the longer timescales considered here (Jamieson *et al*. 2002).

Surface processes

In CS models, the surface processes model specifies the local erosion rate as $\dot{e}\,(t,x)$ = slope × $f(t)$ × $g(x)$, where slope is the local surface slope determined from the Eulerian finite element mesh, $f(t)$ is a time function, and $g(x)$ is a 'climate' function (Fig. 2). To a first approximation $g(x)$ is a measure of the spatial variation of aridity (0 = dry, 1 = wet) across the model. In the CS models described here $f(t)$ is constant, but it varies in the HT-series models (Appendix; Jamieson *et al.* 2004b, this volume). There are no surface processes in the UMS models.

Numerical parameters

For the CS models, the initial dimensions of the Lagrangian crustal grid are 5000 x 35 km (501 x 41 nodes); each element is initially 10 km wide and 0.875 km deep. The Eulerian mechanical grid has 201 x 41 nodes (2000 x 35 km; crust only) and the thermal grid (crust and mantle) has 201 x 68 nodes (2000 x 96 km on undeformed pro-side, Fig.2a). In the diagrams, deformation is displayed using a passive marker grid in which initial vertical markers are spaced at 40 km and horizontal markers at 5 km, with heavy vertical lines initially at 200 km intervals. Model times are quoted either in My (millions of years after start of model) or Ma (millions of years before end of model). The length of the timesteps, Δt, is 3000 y in the CS models and 1000 y in the UMS models.

Crustal scale model results

In this section we describe results from three CS models. Model LHO-1 illustrates Mode 1, Homogeneous Channel Flow, LHO-2 illustrates Mode 2, Heterogeneous Channel Flow, and LHO-3 illustrates Mode 3, Hot Fold Nappes. The models are identical except for the properties of the lower crust, as described below.

Model LHO-1: Homogeneous channel flow

Model LHO-1 is a typical Mode 1 laterally homogeneous model with a uniform 10 km thick lower crust with model rheology *B*(DMD/5)*. This scaling achieves an effective strength that is intermediate between very strong diabase (*B*(DMD))* and intermediate granulite (e.g. Pikwitonei granulite, Mackwell *et al*. 1998, with effective strength *B*(DMD/10))*. The pro- and retro-sides of the model (Figs. 3-6, Parts 1 and 2, respectively) indicate how the two sides of the model evolve with and without surface erosion, respectively. The results are illustrated as pairs of figures in which the first pair shows the material distribution and the deformation of a passive, initially rectangular, Lagrangian marker mesh for the pro- and retro-sides of the model. The bold vertical mesh lines are numbered relative to the surface suture (the initial boundary between the pro- and retro-sides of the model), labeled '0' and located above the model S point. The second pair of figures shows the corresponding distribution of crustal radioactive heating and the temperature field. The third pair shows the distribution of the second invariant of the stress, and the last pair shows the velocity, plotted as vectors, and the second invariant of the strain rate field. Convergence is symmetric with $V_P = 1.5$ cm/y, $V_R = -1.5$ cm/y, and $V_S = 0$.

During the initial 25 My the main style of deformation, shown by the velocity vectors and Lagrangian marker grid, is characterised by diachronous near-pure-shear thickening of the upper- and most of mid-crust, the development of a sub-horizontal shear zone near the base of the mid-crust, and the viscous decoupling of the relatively weak lower mid-crust from the stronger *B*(DMD/5)* lower crust (Figs. 3b & 6b, 20 My). The lower crust is weakly sheared and thickened where the basal boundary condition forces it to detach near the centre of the model. This effect is probably not realistic; Beaumont *et al*. (2004) argued that lower crust is most likely subducted during orogenesis because orogenic antiformal cores comprising thickened lower crust, like that seen here, are not observed in natural orogens. However, lower crust is not subducted in this model to be consistent with the next two models.

The temperature field is closely linked to the evolving distribution of heat-producing material. During diachronous crustal thickening there is some radioactive internal self-heating, but significant thermal disequilibrium remains owing to vertical advection of the temperature field during crustal thickening (Fig. 4b & c, 20 and 30 My). Thermal re-equilibration, by radioactive self-heating and thermal diffusion, occurs with a timescale of close to 20 My during which time the temperature in the lower crust reaches 800°C (Fig. 4b). This self-heating timescale is much shorter than the 50-200 My required for lithospheric-scale thermal relaxation.

At approximately 25 My, channel flow starts in the melt-weakened retro-midcrust, where $T \geq 750°C$, and is soon followed by an equivalent flow in the pro-crust (Figs. 3c & 6c, 30 My). Although the flows develop against the thickened lower crust, this strong antiform does not cause the channel flow by acting as a backstop. Similar channel flows occur in models where the strong core is absent (Beaumont *et al.* 2004; Jamieson *et al*. this volume). The minor asymmetry in the flow results from erosion on the pro-side of the model. The oppositely-directed channel flows subsequently tunnel outward such that their tips evolve with the temperature field, coinciding with the 750°C isotherms (Figs.3, 4 and 6), which are also close to the edges of the orogenic plateau that develops in the centre of the model (Figs. 3 and 4, 30-60 My). Channel flow is restricted to the region beneath the plateau and does not penetrate into the foreland crust, which is too cold. The only significant difference between the two sides is the erosional uplift and exhumation of the pro-flank, which causes tectonic thickening of the mid-crust but is not sufficiently intense to exhume the channel, which continues tunneling. Flow in the channel beneath the plateau reaches velocities of approximately 0.75 m/y; strain rates exceed 10^{-13}/s in the boundary layers but the second invariant of the stress in the channel does not exceed 1 MPa. This model illustrates Homogeneous Channel Flow (Figs 3 and 6).

Model LHO-2: Heterogeneous channel flow

In many orogens the crust of the colliding continents may be heterogeneous. This is almost certainly true in the Himalayan-Tibetan orogen, where the Indian and Asian crusts have different compositions and, moreover, the earlier accretionary history may have given the Asian crust considerable internal heterogeneity. Although we have not undertaken an exhaustive sensitivity analysis of the effects of crustal strength variations, we have a range of model results that include upper, mid- and lower crustal heterogeneities. Model LHO-2 (Figs. 7-10) provides some insight into the effect of variations in lower crustal properties on the thermal-tectonic style of the models. We focus on the relative styles of deformation of the mid- and lower crust and their differences compared with homogeneous lower crust, LHO-1.

The only difference between models LHO-1 and LHO-2 is that the lower crust in the interior of LHO-2 comprises alternating 250 km wide zones with *B*(DMD)* and

B(DMD/10)* rheologies. The external parts of the model have lower crust with *B*(DMD).* The high and low viscosity regions in the lower crust therefore have a nominal viscosity contrast of 10, designed to correspond approximately to the difference between dry, refractory mafic lower crust and intermediate granulite lower crust (e.g. Pikwitonei granulite, see above). However, this factor of 10 contrast is modulated by the nonlinear effect of power-law flow and temperature variations. The strong lower crustal blocks are therefore nominally a factor of 2 stronger than LHO-1 lower crust, and the weak blocks are a factor of 5 weaker.

Model LHO-2 results show a complexly deformed crust that can be understood as the superposition of two main deformation phases. Phase 1 activates and deforms the zones of weaker lower crust in the transition zone between the foreland and the plateau (Fig. 7b). The style is very similar to the deformation of a finite-width salt layer as sediment progrades over it (Lehner 2000); the horizontal pressure gradient in the transition zone between plateau and the foreland acts in the same way as the pressure gradient caused by the prograding sediment (Gemmer *et al.* 2004). It squeezes and evacuates the weak lower crust, then thrusts it and the overlying crust pro-ward on the pro-side, and in the opposite direction on the retro-side, as allochthonous tongues or nappes over adjacent regions of strong lower crust (Fig. 7b-d, 30-50 My). Shears at the leading edges of the tongues propagate upward through the crust, and the allochthonous tongues and their overburden become uplifted and transported. Where lower crust is evacuated it is replaced by subsiding mid-crust, and these regions preferentially shorten and thicken during further contraction (e.g. vertical markers –2 to –3 and -4 to –5, Fig. 7b).

In Phase 2, a channel flow develops in the heterogeneous crust created in phase one. The tongues of overthrust weak lower crust become entrained in the channel flow (Fig. 7d & e, 50-60 My). The remaining zones of strong lower crust are transported into the centre of the plateau and detached at S, where they are incorporated into an antiformal stack (Fig. 7b-e, 30-60 My) similar to that in LHO-1.

The temperature distribution, redistribution of radioactive crust, velocity field, strain rate, and stress in LHO-2 are different from those of LHO-1, but less so than the deformation (Fig. 7) might suggest. Channel flow (Figs. 7 and 10) develops beneath the plateau in both cases. The implication is that heterogeneous lower crust may make the geometry and composition of the channel flows similarly heterogeneous - e.g., in the heterogeneous flow mode the channels may transport detached lumps of much stronger, distinctly different composition, high metamorphic pressure, granulitic or eclogitiic lower crust. Widespread channel flows can develop even under these circumstances, provided that the viscosity of most of the mid-crust becomes sufficiently low and the lumps are not too large to be transported.

Model LHO-3: Hot fold nappes

The evolution of a representative model, LHO-3, designed to test the response of an orogen to collision with successively stronger blocks of lower continental crust is shown in Figures 11-14. The model is symmetric except that one flank of the orogen is mildly denuded by slope-dependent erosion and the other is not. The upper and mid-crust are uniform and the only lateral variation in properties comes from the 250 km long, 15 km thick lower crustal blocks in which the effective power-law viscosity, based on Dry Maryland Diabase *B*(DMD),* is successively reduced by factors of 4, 8, 12, 16, and 20 toward the centre of the model from both sides (Table 1). This scaling creates effective viscosities ranging from *B*(DMD)* in the external crust, through *B*(DMD/10)* (intermediate Pikwitonei granulite, Mackwell *et al*. 1998) to half this value, *B*(DMD/20),* in the centre of the model. The entire lower crust has $\phi^{P}_{eff} = 15°$, but this is not important because deformation occurs in the ductile regime.

The model is highly idealized, and is designed more as a physics/mechanics experiment to test how different strength lower crustal blocks will be absorbed by the model orogen system than as an attempt to model a natural system. In this experiment the blocks that are inserted become progressively stronger with time. The experiment determines when lower crustal blocks appear to be weak, and therefore deform and are incorporated into the orogen or, in contrast, when they are strong and act as indentors. The model represents a development of the vise-type models described by Ellis *et al*. (1998).

The model exhibits a 3-phase evolution. During Phase 1 convergence, the crust containing the weaker lower crustal blocks diachronously shortens and thickens by nearly uniform contraction in the upper and mid-crust (Fig. 11b, Parts 1 &2, 30 My). A ductile shear zone develops at the base of the crust, detaching the overlying weak lower crustal blocks from the basal boundary condition that represents kinematically underthrusting mantle lithosphere (Figs 11b & 14b, 30 My). Diachronous thickening of the radioactive crust is relatively fast and creates thermal disequilibrium owing to the vertical stretching (Fig. 11b, 30 My). This thermal disequilibrium is reduced during Phase 2, a period of radioactive self-heating and thermal relaxation that produces hot, ductile lower crust, highly ductile mid-crust and a relatively cool, strong, frictional-plastic upper crust (Figs 12 & 13, b & c, 30-40 My). Phase 2 is also diachronous and typically takes ca. 20 My after crustal thickening ends (Fig. 12c, 40 My). Phases 1 and 2 occur sequentially as weak lower crustal blocks are inserted, thickened, absorbed, and heated, as the model orogen becomes progressively wider and hotter.

The onset of Phase 3 coincides with the arrival and underthrusting of a lower crustal block that cannot be absorbed by Phase 1-style deformation because it is too strong and resists decoupling. This effect is initially progressive - blocks with rheology based on *B*(DMD/20)*, *B*(DMD/16)* and *B*(DMD/12)* decouple easily and there is no significant change in deformation style. However, the *B*(DMD/4)* block offers some resistance to decoupling, forcing additional contraction on the interior of the system which responds by developing large-scale lower-crustal folds (Fig.11c, 40 My). The transition to Phase 3 becomes fully developed with the arrival of the *B*(DMD)* lower crust. It does not decouple and, consequently, acts as an indentor/plunger that forces weak middle and lower crust into large-scale, gently inclined, ductile fold nappes rooted at the Moho (Fig.11d, 50 My). Some of these are then expelled over the indentor and either inserted into the middle crust (Fig.11e Part 2, 65 My) and/or exhumed to the surface by erosion (Fig.11 Part 1d & e, 50 and 65 My). Surface denudation during Phase 3 determines the relative amount of uplift and exhumation of the fold nappes versus their horizontal transport once inserted into the mid-crust (pro- *vs* retro-sides, Fig. 11). If there is little or no erosion (retro-side), the nappes remain buried and are transported together with the overlying crust, which shows little deformation associated with nappe insertion (Fig.11e Part 2, 65 My). As explained below, the Hot Fold Nappe style of crustal flow is favoured by weak lower crust in the interior of the orogen. The extent of weakening is related to the incubation time, the duration of Phase 2 for each part of the model crust (see Discussion below).

Evolution of topography in LHO models

The evolution of the topography in models LHO-1 to LHO-3 is shown (Fig.15) for two possible isostatic balances in which the density difference between the crust and mantle in either 500 or 600 kg/m^3. The 500 kg/m^3 results are comparable to natural orogens, and predict average plateau elevations of approximately 5500m for models LHO-1 and LHO-2. Model LHO-3 has a higher mean plateau elevation because the strong lower crust is thicker than in other two models. In all three cases the topography has the triangular initial shape expected for small, bivergent, critical wedge type orogens. In LHO-1 and LHO-2 this geometry grows self-similarly during the first 20 My, also as expected for critical wedges, and the maximum elevation reaches 7 to 8 km. After ca. 20 My, radioactive self-heating is sufficient to weaken the mid-crust progressively in the centre of the orogen and the geometry evolves to a central plateau flanked by younger, stronger critical wedges. In LHO-1 and LHO-2 the elevation of the plateau is lower than the local high in the centre of the orogen, which correlates with the antiformal core of strong lower crust (Figs 3 & 7). As noted above, this core does not form if lower crust is subducted . In LHO-2, the plateau evolves less uniformly than in LHO-1 because the topography is sensitive to the absorbtion of the alternating strong and weak lower crustal blocks. LHO-3 differs from the other two models because the lower crustal blocks within the orogen are initially weak, giving low taper angle bivergent critical wedges (e.g Fig 15, 15 My). Later the topography takes the form of a plateau ‘bookended’ by high taper angle strong lower-crustal wedges (e.g. Fig. 15, 37-65 My). The strength of these bounding critical wedges maintains a higher, narrower plateau in LHO-3 compared to the other models.

Upper mantle scale models

The models described in the previous section treat the coupled thermal-mechanical deformation of the crust in a self-consistent manner subject to the assumed basal kinematic velocity boundary conditions. These conditions can be interpreted either as symmetric convergence and subduction of the two mantle lithospheres (Figs 6 ,10 & 14), or as advancing subduction of the pro-mantle lithosphere coupled to the retro-mantle lithosphere (Fig. 2), a style also referred to as ablative subduction (Tao & O'Connell 1992; Pope & Willett 1998). Is this prescribed subduction of the mantle lithosphere dynamically consistent? We describe here two upper mantle scale models (USM) to show that subduction is dynamically consistent for two particular sets of mantle lithosphere properties. An investigation of the range of parameter values for which the models exhibit subduction will be published elsewhere.

Upper mantle scale (UMS) orogenic thermal-mechanical models with viscous-plastic rheologies have been presented by Pysklywec (2001), Pysklywec *et al.* (2000, 2002), and Pysklywec & Beaumont (2004). Rayleigh-Taylor (R-T) instabilities in mantle lithospheres with linear and nonlinear viscosities have been investigated by, for example, Conrad & Molnar (1997), Houseman & Molnar (1997), Molnar *et al.* (1998), and Neil & Houseman (1999). The results of the viscous-plastic experiments demonstrate several modes of mantle lithosphere deformation including subduction, double subduction, and slab breakoff, in addition to the viscous R-T dripping. However, most of this work focused on the early stages of continental-continent collision. Here, our concern is flow modes in large-hot orogens, therefore we describe two examples that illustrate the types of crustal flow and mantle lithosphere behaviour that may occur during prolonged continent-continent collision.

Description of UMS model experiments

Both models are 2D and the domain is 2000 x 600 km (Fig.16). The ALE finite element techniques and the reasons for the choice of model properties are explained in the numerical calculation section and the appendix. The models include the lithosphere and upper mantle (Fig. 16 and Table 2) and are laterally uniform except for a narrow weak zone in the crust and uppermost mantle designed to represent a simplified suture that localizes the initial deformation (Fig. 16). No precursor phase of oceanic subduction is considered in these experiments. The lithosphere boundary conditions are specified at the sides of the model domain and are designed to correspond approximately to the collision of India with Asia. Pro-lithosphere, equivalent to India, converges from the left at a uniform velocity (in both depth and time), V_P=5 cm/y, against a stationary retro-lithosphere, $V_R = 0$ cm/y, at the right boundary corresponding to Asia. In contrast to the CS models described above, velocity is not specified at the base of the crust, nor anywhere inside the model. The subduction advance/retreat velocity, V_S, is instead determined by the dynamical evolution of the model. The sides and base of the model have free-slip boundary conditions and a small, uniform, symmetric, outward leakage flux of material is specified through the sublithospheric mantle parts of the side boundaries to balance the flux of pro-lithosphere into the model (Table 2).

No surface erosion or deposition occurs in these models, which are designed for comparison with the simple tunneling mode of homogeneous channel flow (Beaumont *et al.* 2004, Figs.12a and 13a). This approach was chosen in order to focus on the effect of the mantle lithosphere behaviour on the crustal channel flow.

The mechanical and thermal properties of the UMS models are described in the numerical calculation section and in Table 2; key properties are summarised in Figure 16. In particular, they include frictional-plastic flow, power-law creep, crustal radioactive heating, melt weakening, and the granulite to eclogite phase transition in the lower crust. The crust is similar to that in the CS models, except that the frictional-plastic rheology strain-softens from ϕ_{eff} = 15° to 2° over the range 0.5 to 1.5 of the second invariant of the strain. The mantle lithosphere strain-softens in the same manner. Lower crustal density changes from 2950 to 3100 kg/m^3 during the granulite-eclogite phase transition. This density increase is chosen to be relatively small because only a fraction of the crust is considered to transform to high-density eclogite. The scaled power-law creep parameters for the model layers are given in Figure16 and are discussed above and in the appendix. Density varies among the model layers and with a volume coefficient of thermal expansion of 3 x 10^{-5} /°C. The only difference between the two models is in the reference density of the mantle lithosphere, which is 3300 kg/m^3 in model LHO-LS1 and 3310 kg /m^3 in model LHO-LS2, resulting in a nominal average density difference between the mantle lithosphere and sublithospheric mantle of 40 and 30 kg/m^3, respectively. The results show that the model behaviour is very sensitive to this 0.3% difference in mantle lithosphere density.

Upper mantle scale model results: Models LHO-LS1 and LHO-LS2

Model results are described using the 'pro- retro-' terminology because the dynamic behaviour is similar to the prescribed subduction in the CS models. During the initial stages of convergence in both models (e.g. Fig.17a for LHO-LS1) the mantle lithosphere asymmetrically underthrusts and subducts at a relatively low angle in a 'plate-like' manner with little internal deformation. By 9 My, however, the behaviours of the subducted slabs diverge. The denser mantle lithosphere in LHO-LS2 begins to sink, in addition to subducting, and the lower part of the slab steepens and dips at high angle (Fig. 18a). In contrast, the slab in LHO-LS1 resists subduction and the retro-mantle lithosphere deforms to accommodate the contraction. This subsequently develops into advancing (Fig. 17b) and then double subduction (Fig.17c), during which the subduction point advances dynamically leading to a net subduction zone advance of approximately 700 km between 9 and 33 My, corresponding to an average $V_S = 3.2$ cm/y. At approximately 30 My, the buoyancy of the double slab becomes sufficiently negative that viscous necking starts, leading to breakoff of the double slab at 42 My (Fig.17d), by which time the subduction point has advanced by 900 km at an average velocity of 2.7 cm/y. This style of advancing subduction, at approximately half the overall convergence rate, is effectively the same as that prescribed in model HT1 (Beaumont *et al*. 2004), indicating that the prescribed basal velocities are compatible with a dynamical model with properties like LHO-LS1. The detached lump of mantle lithosphere remains in the model domain and tends to circulate upward because it approaches neutral buoyancy as it heats and thermally expands. In nature an equivalent lump may sink into the lower mantle before it approaches thermal equilibrium.

LHO-LS1 also develops a mid-crustal channel flow similar to those in equivalent CS models where the channel tunnels outward and is not exhumed by erosion (Beaumont *et al.* 2004, Fig.11a). The main difference from the CS models is that the lower crust does not subduct efficiently but instead tends to accumulate near the subduction point (Fig.17b-d). Unlike CS model LHO-1 where the lower crust forms a large antiform, the eclogitic lower crust in LHO-LSI pools at the base of the isostatically depressed crust. This difference occurs because the eclogitic lower crust is denser (3100 *vs.* 2700 kg/m^3) and weaker (*B*(DMD/10) vs. B* (DMD/5)*) than the lower crust in LHO-1.

In contrast to LHO-LS1, the mantle slab in LHO-LS2 is slightly denser and becomes unstable, necks, and breaks off much earlier, between 9 and 12 My (Fig.18b). The slab is sufficiently dense that it subducts without significant deformation of the retro-mantle lithosphere. There is, however, still a significant component of subduction zone advance between 9 and 18 My. Between 18 and 21 My, the subducting slab begins to sink such that its motion is vertically downward along much of its length. Sinking is faster than the overall convergence rate and the system changes to subduction zone retreat, creating a progressively widening region between the slab and the retro-mantle lithosphere that is synchronously filled by the rapid influx of low viscosity, hot (1000-1300°C), sublithospheric mantle (Fig. 18c). This region widens to approximately 200 km by 27 My (Fig.18d). The model therefore displays a combination of subduction zone retreat and mantle delamination. The delamination of the mantle lithosphere from the crust is very efficient because it creates net subduction zone retreat despite the continued convergence of the pro-lithosphere. The delamination velocity therefore exceeds 5 cm/y. The transition to retreating subduction is markedly different from the basal boundary conditions used in the Himalayan-Tibetan crustal scale models (Beaumont *et al*. 2001 & 2004). It has more in common with the behaviour envisaged in the Willett & Beaumont (1994) retreating subduction model, except that the polarity is reversed. Despite the different style of subduction, LHO-LS2 also develops a mid-crustal channel flow (Fig.18c & d), in this case confined to the retro-side of the system. This restriction occurs because delamination and subduction zone retreat occur beneath the converging pro-crust so fast that it does not have time to melt weaken before it is transferred across the migrating subduction point to the retro-side of the system. The overall width of the channel zone is similar to that of LHO-LS1 although the location of subduction beneath the plateau and region of channel flow is different. LHO-LS2 achieves an end-member geometry in which the subducting mantle lithosphere continuously peels away from the crust beneath the leading edge of the plateau and subducts at one side of the orogen. In LHO-LS1 the subduction zone advances beneath the plateau. In the context of the Himalayan-Tibetan system, these two results correspond approximately to subduction of the Indian mantle lithosphere respectively beneath the Indus-Tsangpo and Bangong sutures.

The two UMS models illustrate how sensitive the behaviour of the mantle lithosphere may be to small differences in the density contrast between the mantle lithosphere and the sublithospheric mantle. This sensitivity is enhanced by high temperatures in the lithosphere, which render it weak and prone to changes in the style of subduction. The behaviours seen in these models (and others not reported here) may all occur in nature depending on the ambient conditions during continent-continent collision. The results also indicate that crustal channel flows develop during dynamical subduction that exhibits both subduction zone advance and retreat. This demonstrates that model crustal channel flows are not an artifact of the assumed basal boundary conditions in the CS models. However, the full range of parameter combinations needs to be investigated before drawing additional conclusions.

Discussion

Flow modes in temperature-magnitude space

Can we predict which flow modes will operate in different types of orogens? This question can be answered in a general way using the T-M diagrams (Fig. 1) adapted to show where flow modes are predicted in T-M space (Fig. 19). Mid-crustal flows will not occur in small-cold orogens because typical quartzo-feldspathic crust has a viscosity that is too high at the ambient temperatures. However, orogens that are rich in limestone and evaporite (e.g. calcite, anhydrite, and halite rheologies), which are much weaker than quartz-dominated lithologies, may develop these flows in the small-cold parts of T-M space. Mode boundaries in T-M space (Fig.9) are therefore sensitive to the composition of the lithosphere. For example, the flow modes we have described are common in passive margin salt tectonics (e.g., Lehner 2000; Gemmer *et al*. 2004), despite the small size and cool temperatures of these systems.

For typical quartzo-feldspathic crust, homogeneous channel flows are restricted to the hot regions of T-M space. Their lower limit (Fig. 19) is the threshold at which increasing temperature, which lowers viscosity, and increasing orogen magnitude, which amplifies the gravitational-driving force, combine to drive channel flow. However, because orogen magnitude is limited by the maximum thickness of the crust (ca. 70 km), the range of gravitational force is smaller than that of the variation of viscosity. It follows that the gravitational forcing cannot overcome high crustal viscosity and by implication the crust must be very weak, and presumably hot, for channel flows to occur. Volcanic arcs may represent a small-hot orogen end member that could develop localised channel flows even at relatively small magnitude (Fig. 19).

The tectonically-driven Hot Fold Nappes mode can occur in a much larger part of T-M space, including the homogeneous channel domain (Fig. 19). All that is required is a sufficiently hot and thick orogen interior that nappes will be expelled and injected at the mid-crustal level during indentation. The hot fold nappe behaviour illustrated by model LHO-3 is a variation of that produced by the mechanical vise model, which has been applied to the Newfoundland Appalachians (Ellis *et al*. 1998). Vise-type deformation, where weaker crust is squeezed between 'jaws' of stronger crust, can occur in most of T-M space. Hot fold nappes (Fig. 19), however, form only when the upper parts of the vise jaws are sufficiently weak that they cannot resist expulsion of the nappes when weak material is expelled from the interior over the vise or indentor. Jamieson *et al*. (2004a) have interpreted part of the Grenville orogen to record the diachronous evolution of hot fold nappes, possibly with superimposed heterogeneous channel flows, during Mesoproterozoic collision on the Laurentian margin. We suspect that similar tectonically driven styles will be recognized in many North American orogens, which developed by successive collisions against and accretion to the cratonic core which acts as an indentor. In particular, the Trans Hudson, central and southern Appalachian, and southern Canadian Cordilleran orogens, and the Archean Slave and Superior cratons, are prime candidates for these flows.

The Heterogeneous Channel Flow mode is transitional between the other flow modes described here and has both tectonic forcing, required to activate and evacuate weak lower crust, and gravitational forcing required for channel flow. The fully developed form of this flow mode therefore overlaps with the homogeneous channel flow region of T-M space because both require gravitationally-driven flow (Fig. 19). The tectonic evacuation of weak lower crust alone can, however, occur in a larger region of T-M space, which grades into the hot fold nappes domain (Fig. 19).

In nature, unlike our 2D models, ductile flow will be not be restricted to the direction of convergence. In many situations, flow (sub-)parallel to the strike of the orogen may be preferred over expulsion of hot nappes, and channel flow may also be directed around strong enclaves of crust. The inferred outward movement of crust beneath the eastern flank of the Tibetan plateau (e.g. Clark & Royden 2000) is one example of a three-dimensional flow, and Hatcher & Merschat (this volume) describe evidence for Paleozoic orogen-parallel flows in the southern Appalachians.

Effect of thermal relaxation and incubation time on crustal flows

Comparing the timescales of external and internal orogenic processes helps to predict the flow styles in the model experiments. We define an external timescale, the incubation time, to be the lag time between rapid tectonic thickening of the crust during contractional orogenesis and a subsequent external process, such as indentation, that acts on the system. This definition corresponds to that of England & Thompson (1984) in which erosion was the external process. Note that the incubation time varies with position within the model because tectonic thickening is diachronous. The orogen response to indentation will depend on whether the incubation time is long or short compared with timescales of internal processes, defined below, that are required for radioactive heating and thermal relaxation to achieve particular thermal-rheological threshold states within the crust.

In the case of the hot fold nappes mode, let τ_{HN} be the delay time necessary since rapid tectonic crustal thickening for self heating and thermal relaxation to achieve the thermal-rheological state required for the hot nappe type of response seen in LHO-3. That is, hot nappes will be the flow style in models like LHO-3 when the incubation time is approximately equal to τ_{HN}. For shorter or longer incubation times the response may be quite different. Although we have not measured τ_{HN} accurately for the models described here, it is related to the thermal relaxation timescale and is estimated to equal the 20-30 My required for radioactive self-heating to heat the thickened mid-crust to $T \geq$ 700°C (Figs 4b-c,8b & 12b, see also Medvedev & Beaumont, this volume), when ductile nappes will easily form during indentation. Although related, the thermal relaxation timescale and τ_{HN} differ in that the former is the characteristic timescale that measures decay of thermal disequilibrium, whereas the latter is the time required to achieve a particular thermal-rheological state. Tectonic thickening of cold crust with a low level of radioactivity will be associated with a characteristic thermal relaxation timescale, but the crust may never become weak enough for the hot nappe response, giving the system a finite thermal relaxation timescale but an infinite $\tau_{HN.}$ Such orogens are subcritical with respect to any of the three flow modes described here (Fig. 19). Conversely, crust that is already hot and weak before thickening can have a τ_{HN} that is zero or only a fraction of the thermal relaxation timescale. Model LHO-3, which creates hot fold nappes, satisfies the condition that the incubation time, which varies from 20 to 50 My with lateral position at the time of the onset of indentation at 55 My (Fig.11), was equal to or greater than τ_{HN} (20-30 My). Earlier indentation, at 30 My for example, may just have caused more crustal thickening.

Model LHO-1 illustrates the analogous situation in regard to the onset of channel flows. We define τ_{CF} as the delay, since the time of rapid crustal thickening, required for the onset of gravitationally-driven channel flow. Given that channel flow requires weaker crust than is necessary for the development of hot nappes during indentation, τ_{CF} is normally larger than τ_{HN} for the same material properties. This relationship explains LHO-3 behaviour, in which the initial response to indentation is to produce fold nappes (Fig. 11d), because the incubation time is greater than τ_{HN} but less than τ_{CF}. Later, at 65 My, a more pronounced channel flow is superimposed on the system (Fig. 11e) indicating that the incubation time is now greater than τ_{CF}.

Model LHO-2 also illustrates the effects of incubation. During Phase 1 deformation, the weaker lower crustal regions are detached and evacuated during their incorporation into the model orogen (Fig. 7b-d). These regions are already sufficiently hot and weak that they do not require incubation in order to form hot nappes. In contrast, the overlying mid-crust initially deforms by shortening and thickening (Fig. 7b). The gravitationally driven channel flow (Fig. 7c-e) develops only during the diachronous second phase, and has a τ_{CF} of approximately 30 My.

In summary, comparison of the thermal incubation time with τ_{HN} and τ_{CF} provides a guide to the flow modes that will develop in the models. Diachroneity of crustal tectonic thickening implies that different flow modes can coexist in different regions of the model.

For natural orogens we may know the timing of tectonic thickening of the crust but lack information on the thermal-rheological evolution. In the absence of good estimates of τ_{HN} and τ_{CF} indirect evidence of crustal viscosity and temperature can be derived from large-scale topography and magmatism. Plateau development in collisional orogens demonstrates that the crust, and possibly the entire lithosphere, cannot sustain thickness variations against gravitational forces and has flowed to equilibrate the pressure in the crust below the plateau to a lithostatic state. This state indicates the crust is weak and that it is more likely to flow over a lower crustal indentor than to shorten and thicken against it. The development of a plateau is therefore a measure of a thermo-rheologically weak crust, and the delay from crustal thickening to initial plateau development is an approximate estimate of τ_{HN} and a lower bound on $\tau_{CF.}$ Magmatism, particularly involving widespread crustal melts, is an independent indicator that crustal temperature is high and that the crust is therefore probably weak. The delay from orogenic crustal thickening to the onset of magmatism can also be used as an upper bound estimate of τ_{HN}, and for crustal melts also provides an estimate of $\tau_{CF.}$

Infrastructure and Superstructure

All three LHO models illustrate the development of differing styles of deformation at different levels of the crust corresponding to what is termed 'Infrastructure and Superstructure' in the classical geological literature. The terminology refers to contrasting styles of deformation and metamorphism in the upper, superstructure, and lower, infrastructure, levels of the crust. The superstructure preserves early, low-grade deformation, typically with upright contractional structures, whereas the infrastructure is a ductile, high-grade, migmatitic level with younger gently dipping structures that overprint early structures (Culshaw *et al*. 2004); see also Williams *et al*. this volume. In the LHO models the superstructure develops during crustal shortening and thickening and the infrastructure is superimposed by the flows that develop later in the mid- and lower crust. The main differences among the models are in the cause and timing of infrastructure development.

In model LHO-3 (Fig. 11), the infrastructure comprises lower and mid-crustal nappes partly overlying the underthrust indentor and decoupled from the upper crust by a reverse-sense shear zone at the top of the highly strained ductile mid-crust (Fig. 11,d & e). The upper crustal superstructure remains relatively undeformed after Phase 1 shortening except where exhumed by syntectonic erosion (pro- *vs.* retro-sides of Fig. 11d & e). In contrast, Phase 1 structures in the mid- and lower-crustal infrastructure are strongly overprinted by Phase 3 flow (Fig. 11d & e). From an observational perspective, the three-phase evolution of model LHO-3 leads to what would be recognized geologically as an old, but not reworked, contractional upper crust underlain by and decoupled from mid- and lower-crust that records the initial contraction, thermal relaxation, and the superimposed deformation activated by Phase 3 collision with the indentor. The mid- and lower-crust become weak, with τ_{HN} approximately 20-30 My, but hot-nappe deformation is not strongly activated until the indentor collides much later. In this example the infrastructure is created by the tectonic indentation process, not by internal gravitational flow, after a long incubation time. In contrast, the infrastructure in models LHO-1 and LHO-2 becomes sufficiently weak during thermal relaxation that it deforms and flows under gravitational forces alone (Figs. 3 and 7). Under these circumstances the development of superstructure/infrastructure relationships in LHO-1 is governed by the delay timecale τ_{CF} for channel flows and does not require external forcing by an indentor or some other tectonic process. In LHO-2 the formation of the lower crustal infrastructure starts during the initial crustal shortening and thickening and is subsequently overprinted by the channel flow on the τ_{CF} timescale.

Conclusions

Numerical models have been used to investigate crustal flows in large-hot orogens in plane strain at the crustal and upper mantle scale. The flow styles are divided into three types, Homogeneous, Heterogeneous, and Hot Fold Nappe modes, and the conditions under which each will operate are assessed in the framework of the orogenic T-M diagram, which plots orogen characteristics in terms of the two principal controls, temperature and magnitude. We draw five main conclusions.

1) Gravitationally driven mid-crustal channel flows, exemplified by the Homogeneous and Heterogeneous modes (Fig. 20), are most likely to occur in Giant and Supergiant members of the Large Hot Orogen family (Figs 1 & 19),

such as the Himalayan-Tibetan system. Gravitationally driven flows require orogenic crust to be weak (low viscosity) and therefore hot under most circumstances. These flows are facilitated by gravitational forces that result from the large difference in potential energy between the tectonically thickened interior and normal thickness exterior crust in large hot orogens. These flows may have been more prevalent in Archean orogens, if they were indeed hotter than equivalent-sized contemporary orogens.

2) Tectonically forced flow modes, exemplified by the Hot Fold Nappes mode and the tectonic component of the Heterogeneous Flow mode (Fig. 20), may occur in Giant and Supergiant orogens (Figs 1& 19). More importantly, they can occur in large hot orogens that are not hot and/or large enough to undergo gravitationally driven flows. In particular, the Hot Fold Nappes mode is predicted for accretionary and collisional orogens where the orogen experiences late stage collision/indentation by strong crust, for example, older refractory crust such as a cratonic nucleus, or cold oceanic crust.

3) Both crustal scale (CS) models with kinematic mantle subduction basal boundary conditions and upper mantle scale (UMS) models can develop homogeneous crustal channel flows. To a first approximation the general characteristics of these flows are insensitive to the effects of advancing or retreating subduction of the underlying mantle lithosphere. However, a moving subduction zone will change its position relative to the crustal channel and its final position may vary from a location beneath the edge of the plateau (Fig. 17) to one beneath the centre of the plateau (Fig. 18).

4) The Temperature-Magnitude (T-M) diagram, which we have introduced here, provides a framework for the classification of orogens, ranging from small-cold to large-hot, and for their development with respect to a main sequence (Fig. 1). We suggest that natural orogens can be analysed, at least conceptually, by their evolution in T-M space and that this approach offers a way to predict when different types of flow modes may occur.

5) These flow modes have been inferred to have developed in some ancient orogens, including the Mesoproterozoic western Grenville orogen (Jamieson *et al.* 2004a), and the Paleozoic Inner Piedmont of the southern Appalachians (Hatcher & Merschat, this volume). In North America, which has grown outward by successive collisions against and accretion to the cratonic core, which could have acted as an indentor, we anticipate that the tectonically driven styles will be recognized in many large-hot orogens, including the Trans-Hudson, and southern Canadian Cordillera orogens, and parts of the Archean Slave and Superior cratons.

Acknowledgments This research was funded by NSERC Discovery Grants to Beaumont and Jamieson. Beaumont also acknowledges support from the Canadian Institute for Advanced Research, the Canadian Research Chairs programme, and from an IBM Shared University Research grant. The models used software developed by P. Fullsack, Dalhousie University. The work benefited from discussions with colleagues including N. Culshaw, D.Grujic and especially the late D.Nelson. We thank the reviewers G. Houseman and J. Platt for thorough reviews and R. Law for editorial advice. We are grateful to the conference organizers for providing a stimulating environment in which to respond to M. Harrison, and to D.Nelson for originally encouraging us to tackle this problem.

References

Amidon, W.H., Burbank, D.W. & Gehrels, G.E. 2005. U-Pb zircon ages as a sediment mixing tracer in the Nepal Himalaya. *Earth and Planetary Science Letters*, **235**, 244-260.

Beaumont, C. & Quinlan, G. 1994. A geodynamic framework for interpreting crustal scale seismic reflectivity patterns in compressional orogens. *Geophysical Journal International***,** **116***,* 754-783.

Beaumont, C., Fullsack, P. & Hamilton, J. 1994. Styles of crustal deformation caused by subduction of the underlying mantle. *Tectonophysics***,** **232***,* 119-132.

Beaumont, C., Fullsack, P. & Hamilton, J. 1992. Erosional control of active compressional orogens. In: McClay, K.R. (ed) *Thrust Tectonics*, Chapman and Hall, 1-18.

Beaumont, C., Ellis, S., Hamilton, J. & Fullsack, P. 1996a. Mechanical model for subduction - collision tectonics of Alpine-type compressional orogens. *Geology*, **24**, 675-678.

Beaumont, C., Kamp, P.J.J., Hamilton, J. & Fullsack, P. 1996b. The continental collision zone, South Island, New Zealand; Comparison of geodynamical models and observations. *Journal of Geophysical Research,* **101**, 3333-3359.

Beaumont, C., Ellis, S. & Pfiffner, A. 1999. Dynamics of sediment subduction-accretion at convergent margins: short term modes, long-term deformation, and tectonic implications. *Journal of Geophysical Research*, **104**, 17573-17601.

Beaumont, C., Muñoz, J.A., Hamilton, J. & Fullsack, P. 2000. Factors controlling the Alpine evolution of the central Pyrenees inferred from a comparison of observations and geodynamical models. *Journal of Geophysical Research*, **105**, 8121-8145.

Beaumont, C., Jamieson, R.A., Nguyen, M.H. & Lee, B. 2001. Himalayan tectonics explained by extrusion of a low-viscosity crustal channel coupled to focused surface denudation. *Nature*, **414**, 738-742.

Beaumont, C., Jamieson, R.A., Nguyen, M.H. & Medvedev, S. 2004. Crustal channel flows: 1. numerical models with applications to the tectonics of the Himalayan-Tibet orogen. *Journal of Geophysical Research*, **109,** *B06406 doi: 10.1029/2003JB002809.*

Bird, P. 1991. Lateral extrusion of lower crust from under high topography, in the isostatic limit. *Journal of Geophysical Research*, **96**, 10275-10286.

Braun, J. & Beaumont, C. 1995. Three dimensional numerical experiments of strain partitioning at oblique plate boundaries: implications for contrasting tectonic styles in California and South Island, New Zealand. *Journal of Geophysical Research*, **100**, 18,059-18,074.

Chopra, P.N. & Paterson, M.S. 1984. The role of water in the deformation of dunite. *Journal of Geophysical Research*, **89**, 7861-7876.

Clark, M.K. & Royden, L.H. 2000. Topographic ooze: Building the eastern margin of Tibet by lower crustal flow. *Geology*, **28**, 703-706.

Conrad, C.P. & Molnar, P. 1997. The growth of Rayleigh-Taylor-type instabilities in the lithosphere for various rheological and density structures. *Geophysical Journal International*, **129**, 95-112.

Culshaw, N.G., Beaumont, C. & Jamieson, R.A. 2004. Geodynamic models of contrasting structural styles and ages in upper and lower crust of collisional orogens: resolution of the structural vs seismic paradox. *Geological Association of Canada-Mineralogical Association of Canada, Annual Meeting, St Catherine's, Ontario,* *Abstracts*, **29**, 331.

DeCelles, P.G., Gehrels, G.E., Najman, Y., Martin, A.J., Carter, A. and Garzanti, E. 2004. Detrital geochronology and geochemistry of Cretaceous-Early Miocene strata of Nepal: implications for timing and diachroneity of initial Himalayan orogenesis. *Earth and Planetary Science Letters,* **227**, 313-330.

Ellis, S. & Beaumont, C. 1999. Models of convergent boundary tectonics: implications for the interpretation of Lithoprobe data. *Canadian Journal of Earth Sciences*, **36**, 1711-1741.

Ellis, S., Beaumont, C., Jamieson, R. & Quinlan, G. 1998. Continental collision including a weak zone - the vise model and its application to the Newfoundland Appalachians. *Canadian Journal of Earth Sciences*, **35**, 1323-1346.

Ellis, S., Beaumont, C. & Pfiffner, A. 1999. Geodynamic models of crustal-scale episodic tectonic accretion and underplating in subduction zones. *Journal of Geophysical Research*, **104**, 15169-15190.

England, P.C. & Thompson, A.B. 1984. Pressure-Temperature-Time paths in regional metamorphism I. Heat transfer during the evolution of regions of thickened continental crust. *Journal of Petrology*, **25,** 894-928.

Fullsack, P. 1995. An arbitrary Lagrangian-Eulerian formulation for creeping flows and its applications in tectonic models. *Geophysical Journal International*, **120**, 1-23.

Gemmer, L., Ings, S.J., Medvedev, S. & Beaumont, C. 2004. Salt tectonics driven by differential sediment loading: stability analysis and finite-element experiments. *Basin Research,* **16**, 199-208, doi: 10.1111/j.1365-2117.2004.00229.x

Gleason, G.C. & Tullis, J. 1995. A flow law for dislocation creep of quartz aggregates with the molten salt cell. *Tectonophysics*, **247**, 1-23.

Grujic, D., Casey, M., Davidson, C., Hollister, L.S., Kundig, R., Pavlis, T. & Schmid, S. 1996. Ductile extrusion of the Higher Himalayan Crystaline in Bhutan: Evidence from quartz microfabrics. *Tectonophysics*, **260**, 21-43.

Grujic, D., Hollister, L.S. & Parrish, R.R. 2002. Himalayan metamorphic sequence as an orogenic channel: Insight from Bhutan. *Earth and Planetary Science Letters*, **198**, 177-191.

Hatcher, R.D. & Merschat, A.J. this volume. The Appalachian Inner Piedmont: an exhumed strike-parallel, tectonically forced orogenic channel. *In*: Law, R.D., Searle, M.P. & Godin, L. (eds) *Channel Flow, Extrusion, and Exhumation in Continental Collision Zones.* Geological Society, London, Special Publications, .

Hertszprung, E. 1905. “Zure Strahlung der Stern,” Zeitschrift für wissenschaftliche Photographie 3, 429-442. The discovery, from the spectra, that some stars are giants and others dwarfs. A translation, with commentary, appears in Lang, Kenneth R. & Owen Gingerich (eds), *A Source Book in Astronomy & Astrophysics*, 1900-1975 (Harvard Univ. Press, Cambridge, MA, 1979), 208-11.

Hirth, G., Teyssier, C. & Dunlap, W.J. 2001. An evaluation of quartzite flow laws based on comparisons between experimentally and naturally deformed rocks. *International Journal of Earth Science* (Geologische Rundschau), **90**, 77-87.

Hodges, K.V. 2000. Tectonics of the Himalaya and southern Tibet from two perspectives. *Geological Society of America Bulletin*, **112**, 324-350.

Houseman, G.A. & Molnar, P. 1997. Gravitational (Rayleigh-Taylor) instability of a layer with non-linear viscosity and convective thinning of continental lithosphere.

Geophysical Journal International, **128**, 125-150.

Huerta, A.D., Royden, L.H. & Hodges, K.V. 1998. The thermal structure of collisional orogens as a response to accretion, erosion, and radiogenic heating. *Journal of Geophysical Research*, **103**, 15287-15302.

Huismans, R.S. & Beaumont, C. 2003. Symmetric and asymmetric lithosphere extension: relative effects of frictional-plastic and viscous strain softening, *Journal of Geophysical Research*, **108** (B10), 2496, doi:10.1029/2002JB002026.

Jamieson, R.A., Beaumont, C., Fullsack, P. & Lee, B. 1998. Barrovian regional metamorphism: Where's the heat? In: Treloar, P. & O'Brien, P. (eds) *What Controls Metamorphism and Metamorphic Reactions?* Geological Society, London, Special Publications, **138**, 23-51.

Jamieson, R.A., Beaumont, C., Nguyen, M.H. & Lee, B. 2002. Interaction of metamorphism, deformation, and exhumation in large convergent orogens. *Journal of Metamorphic Geology*, **20**, 9-24.

Jamieson, R.A., Culshaw, N., Beaumont, C., Nguyen, M.H. & Slagstad, T. 2004a. Hot nappes and lumpy channels: mid-crustal flow modes in the western Grenville orogen. In: *Channel Flow, Ductile Extrusion and Exhumation of Lower-mid Crust in Continental Collision Zones* Abstracts with Programme. Geological Society of London.

Jamieson, R.A., Beaumont, C., Medvedev, S. & Nguyen, M.H. 2004b. Crustal channel flows: 2. numerical models with implications for metamorphism in the Himalayan-Tibetan orogen. *Journal of Geophysical Research,* **109***, B06407,* doi:10.1029/2003JB002811.

Jamieson, R.A., Beaumont, C., Nguyen, M.H. & Grujic, D. (this volume) Provenance of the Greater Himalayan Sequence and associated rocks: Predictions of channel flow models. *In:* Law, R.D., Searle, M.P. & Godin, L. (eds) *Channel flow, ductile extrusion, and exhumation in continental collision zones.* Geological Society, London, Special Publications

Karato, S. & Wu, P. 1993. Rheology of the upper mantle: a synthesis, *Science*, **620**, 771-778.

Karato, S. & Jung, H. 2003. Effects of pressure on high-temperature dislocation creep in olivine. *Philosophical Magazine*, **83**, 404-414.

Klemperer, S.L. 2005. (this volume) Crustal flow in Tibet: A review of geophysical evidence for the physical state of Tibetan lithosphere. *In:* Law, R.D., Searle, M.P. & Godin, L.. (eds) *Channel flow, ductile extrusion, and exhumation in continental collision zones.* Geological Society, London, Special Publications

Lehner, F.K. 2000. Approximate theory of substratum creep and associated overburden deformation in salt basins and deltas. In: Lehner, F.K. & Urai, J.L. (eds) *Aspects of Tectonic Faulting*, Springer-Verlag, Berlin, 21-47.

Mackwell, S.J., Zimmerman, M.E. & Kohlstedt, D.L. 1998. High-temperature deformation of dry diabase with application to tectonics on Venus. *Journal of Geophysical Research*, **103**, 975-984.

Medvedev, S. & Beaumont, C. this volume. Growth of continental plateaus by crustal channel injection: Constraints and thermo-mechanical consistency *In*: Law, R.D., Searle, M.P. & Godin, L. (eds) *Channel Flow, Extrusion, and Exhumation in Continental Collision Zones.* Geological Society, London, Special Publications, Molnar, P., Houseman, G.A. & Conrad, C.P. 1998. Rayleigh-Taylor instability and convective thinning of mechanically thickened lithosphere: effects of non-linear viscosity decreasing exponentially with depth and of horizontal shortening of the layer. *Geophysical Journal International*, **133**, 568-584.

Najman, Y. 2006. The detrital record of orogenesis: A review of approaches and techniques used in Himalayan sedimentary basins. *Earth Science Reviews*, **74**, 1-72.

Najman, Y., Carter, A., Oliver, G. & Garzanti, E. 2005. Provenance of Eocene foreland basin sediments, Nepal: Constraints to the timing and diachroneity of early Himalayan orogenesis. *Geology*, **33**, 309-312.

Neil, E.A. & Houseman, G.A. 1999. Rayleigh-Taylor instability of the upper mantle and its role in intraplate orogeny. *Geophysical Journal International*, **138**, 89-107.

Nelson, K.D., Zhao, W., Brown, L.D., Kuo, J., Che,J., Liu, X., Klemperer,S.L., Makovsky, Y., Meissner, R., Mechie, J., Kind, R.,Wenzel, F., Ni, J., Nabelek, J., Chen, L., Tan, H., Wei, W., Jones A.G., Booker, J., Unsworth, M., Kidd, W.S.F., Hauck, M., Alsdorf, D., Ross A., Cogan, M., Wu, C., E.A. Sandvol, E.A. & Edwards, M. 1996. Partially molten middle crust beneath southern Tibet: a synthesis of Project INDEPTH results. *Science*, **274**, 1684-1688.

Owens, T.J. & Zandt, G. 1997. Implications of crustal property variations for models of Tibetan plateau evolution. *Nature*, **387**, 37-43.

Pfiffner, O.A., Ellis, S. & Beaumont, C. 2000. Collision tectonics in the Swiss Alps: implications from geodynamic models. *Tectonics,* **19**, 1065-1094.

Pope, D.C. & Willett, S.D. 1998. Thermal-mechanical model for crustal thickening in the

Central Andes driven by ablative subduction. *Geology*, **26**, 511-514.

Pysklywec, R.N. 2001. Evolution of subducting mantle lithosphere at a continental plate boundary. *Geophysical Research Letters*, **28**, 4399-4402.

Pysklywec, R.N. & Beaumont, C. 2004. Intraplate tectonics: feedback between radioactive thermal weakening and crustal deformation driven by mantle lithosphere instabilities. *Earth and Planetary Science Letters*, **221**, 275-292.

Pysklywec, R.N., Beaumont, C. & Fullsack, P. 2000. Modeling the behaviour of the continental mantle lithosphere during plate convergence. *Geology*, **28**, 655-658.

Pysklywec, R.N., Beaumont, C. & Fullsack, P. 2002. Lithospheric deformation during the early stages of continental collision: numerical experiments and comparison with South Island, New Zealand. *Journal of Geophysical Research,* **107**, 10129, ETG 3(1-19).

Royden, L.H. 1996. Coupling and decoupling of crust and mantle in convergent orogens: Implications for strain partitioning in the crust. *Journal of Geophysical Research*, **101**, 17679-17705.

Royden, L.H., Burchfiel, B.C., King, R.W., Chen, Z., Shen, F. & Liu, Y. 1997. Surface deformation and lower crustal flow in eastern Tibet. *Science*, **276**, 788-790.

Rosenberg, C.L. & Handy, M.R. 2005. Experimental deformation of partially melted granite revisited: implications for the continental crust*. Journal of Metamorphic Geology,* **23**, 19-28.

Russell, H. N. 1914. Relations between the Spectra and Other Characteristics of the Stars, *Popular Astronomy* **22**, 275-294, 331-351. [excerpted with commentary in Lang, Kenneth R. & Owen Gingerich (eds), *A Source Book in Astronomy &*

Astrophysics, 1900-1975 (Harvard Univ. Press, Cambridge, MA, 1979), 212-20 and also in Shapley, Harlow (ed) *Source Book in Astronomy, 1900-1950* (Harvard University Press, Cambridge, MA, 1960), 253-62.

Shen, F., Royden, L.H. & Burchfiel, B.C. 2001. Large-scale crustal deformation of the Tibetan Plateau. *Journal of Geophysical Research*, **106**, 6793-6816.

Tao, W.C. & O'Connell R.J. 2003. Ablative subduction: A two-sided alternative to the conventional subduction model. *Journal of Geophysical Research*, **97**, 8877-8904.

Turcotte, D.L. & Schubert, G. 1982. *Geodynamics: Applications of Continuum Physics to Geological Problems.* John Wiley & Sons, 450 pp.

Unsworth, M.J., Jones, A.G., Wei, W., Marquis, G., Gokarn, S.G., Spratt, J.E. & the INDEPTH-MT team 2005. Crustal rheology of the Himalaya and Southern Tibet inferred from magnetotelluric data. *Nature*, **438**, 78-81.

Vanderhaeghe, O., Medvedev, S., Fullsack, P., Beaumont, C. & Jamieson, R.A. 2003. Dynamic evolution of orogenic wedges and continental plateaus: insights from thermal-mechanical modeling of convergent orogens. *Geophysical Journal International*, **153***,* 27-51.

Waschbusch, P., Batt, G. & Beaumont, C. 1998. Subduction zone retreat and recent tectonics of the south island of New Zealand. *Tectonics,* **17***,* 267-284.

Westaway, R. 1995. Crustal volume balance during the India-Eurasia collision and altitude of the Tibetan plateau: A working hypothesis. *Journal of Geophysical Research*, **100**, 15173-15192.

Wilks, K.R. & Carter, N.L. 1990. Rheology of some continental crustal rocks. Tectonophysics, 182, 57-77.

Willett, S.D. 1999. Orogeny and orography: The effects of erosion on the structure of mountain belts. *Journal of Geophysical Research*, **104**, 28957-28981.

Willett, S.D. & Beaumont, C. 1994. Subduction of Asian lithospheric mantle beneath Tibet inferred from models of continental collision. *Nature*, **369**, 642-645.

Willett, S., Beaumont, C. & Fullsack, P. 1993. A mechanical model for the tectonics of doubly-vergent compressional orogens. *Geology,* **21**, 371-374.

Williams, P.F., Jiang, D. & Lin, S. this volume. Interpretation of deformation fabrics of infrastructure zone rocks in the context of channel flow and other tectonic models. *In*: Law, R.D., Searle, M.P. & Godin, L. (eds) *Channel Flow, Extrusion, and Exhumation in Continental Collision Zones.* Geological Society, London, Special Publications,

Figure Captions

Fig.1. Orogen Temperature–Magnitude (T-M) diagram. (**a**) Classification of orogen types and comparison of T-M diagram with the Hertzsprung-Russell (H-R) diagram for stars. (**b**) Suggested classification of particular orogens. (**c**) Classification of types of mechanical and thermal-mechanical models (see text) that have been used to model different orogen types according to position in T-M space.(Colour figures and animations available at http://geodynamics.oceanography.dal.ca/LHO/flowmodes/extras.html)

Fig.2. Initial crustal model conditions. Only the central part of the 2000 km long model is shown. (**a**) Passive Lagrangian marker grid and mechanical layers; '0' model surface suture position above subduction point, S. (**b**) Initial thermal structure, radioactive layers, A_1 and A_2 and conductive steady state isotherms, and general velocity vectors, showing convergence with $V_P = 1.5$, $V_R = -1.5$ and $V_S = 0$ cm/y and implied double subduction of the mantle lithospheres beneath S. (**c**) Relationship between initial mechanical and thermal layers and summary of parameters (see also Table 1); effect of reduction in viscosity for quartz-rich upper and mid-crust from flow law value at 700°C to10^{19} Pa.s at 750°C (melt weakening); effective viscosity used in model shown by solid line.

Fig.3. Model LHO-1 results showing evolution of crustal-scale deformation for (**Part 1**) pro crust and (**Part 2**) retro-crust. At the crustal scale the two sides differ only in that the pro-side has surface erosion (V = H). Panels show Lagrangian marker grid and upper, mid- and lower crustal materials. t = elapsed model time (My =million years), Δx = total convergence.

Fig.4. Model LHO-1 results showing evolution of the temperature (isotherms shown at 100°C intervals) and distribution of the crustal radioactive heat production (grey and white areas) for (**Part1**) pro-crust and (**Part 2**) retro-crust; grey (V=H). Heavy line with dots is the position of the suture below the surface marker '0'. t = elapsed model time, Δx = total convergence.

Fig.5. Model LHO-1 results showing evolution of the stress field (second invariant of the stress) for (**Part 1**) pro-crust and (**Part 2**) retro-crust (V=H). Heavy line with dots is

the position of the suture below the surface marker '0'. t = elapsed model time, Δx = total convergence.

Fig.6. Model LHO-1 results showing evolution of the strain rate field (second invariant of the strain rate) and velocity field (horizontal lines are velocity vectors) for (**Part 1**) pro-crust and (**Part 2**) retro-crust (V=H). Heavy line with dots is the position of the suture below the surface marker '0'. t = elapsed model time, Δx = total convergence.

Fig.7. Model LHO-2 results showing evolution of crustal-scale deformation for (**Part 1**) pro crust and (**Part 2**) retro-crust. At the crustal scale the two sides differ only in that the pro-side has surface erosion (V = H). Panels show Lagrangian marker grid and upper, mid- and lower crustal materials. t = elapsed model time, Δx = total convergence.

Fig.8. Model LHO-2 results showing evolution of the temperature (isotherms shown at 100°C intervals) and distribution of the crustal radioactive heat production (grey and white areas) for (**Part 1**) pro-crust and (**Part 2**) retro-crust (V=H). Heavy line with dots is the position of the suture below the surface marker '0'. t = elapsed model time, Δx = total convergence.

Fig.9. Model LHO-2 results showing evolution of the stress field (second invariant of the stress) for (**Part 1**) pro-crust and (**Part 2**) retro-crust (V=H). Heavy line with dots is the position of the suture below the surface marker '0'. t = elapsed model time, Δx = total convergence.

Fig.10. Model LHO-2 results showing evolution of the strain rate field (second invariant of the strain rate) and velocity field (horizontal lines are velocity vectors) for (**Part 1**) pro-crust and (**Part 2**) retro-crust (V=H). Heavy line with dots is the position of the suture below the surface marker '0'. t = elapsed model time, Δx = total convergence.

Fig.11. Model LHO-3 results showing evolution of crustal-scale deformation for (**Part 1**) pro crust and (**Part 2**) retro-crust. At the crustal scale the two sides differ only in that the pro-side has surface erosion (V = H). Panels show Lagrangian marker grid and upper, mid- and lower crustal materials. t = elapsed model time, Δx = total convergence.

Fig.12. Model LHO-3 results showing evolution of the temperature (isotherms shown at 100°C intervals) and distribution of the crustal radioactive heat production (grey and white areas) for (**Part 1**) pro-crust and (**Part 2**) retro-crust (V=H). Heavy line with dots is

the position of the suture below the surface marker '0'. t = elapsed model time, Δx = total convergence.

Fig.13. Model LHO-3 results showing evolution of the stress field (second invariant of the stress) for (**a**) pro-crust and (**b**) retro-crust (V=H). Heavy line with dots is the position of the suture below the surface marker '0'. t = elapsed model time, Δx = total convergence.

Fig.14. Model LHO-3 results showing evolution of the strain rate field (second invariant of the strain rate) and velocity field (horizontal lines are velocity vectors) for (**Part 1**) pro-crust and (**Part 2**) retro-crust (V=H). Heavy line with dots is the position of the suture below the surface marker '0'. t = elapsed model time, Δx = total convergence.

Fig.15. Evolution of the topography for models LHO-1 to LHO-3 shown with respect to the S-point, x=0 (My =millions of years). The height scale is shown for two representative isostatic balances which depend on $\Phi = \Delta\rho/\rho_m$, where $\Delta\rho = \rho_m - \rho_c$ and ρ_m and ρ_c are the mantle and crustal densities, respectively. The scale of the height is most sensitive to $\Delta\rho$ and the results are therefore shown for $\Delta\rho$ = 500 and 600kg/m^3, corresponding to Φ=0.156 and 0.182, respectively.

Fig.16. Configuration and principal properties of the upper mantle scale models, LHO-LS1 and LHO-LS2 (see also Table 2). These models are the same except for the minor difference in the reference densities of their respective mantle lithospheres. Notation $\phi = 15^\circ \rightarrow 2^\circ$ implies strain softening of the internal angle of friction in this case over the range of strain of the second invariant of *0.5* to *1.5.* Effective viscosity η: *B*(WQ× 5)* (Wet Quartz rheology, scaled by 5); *B*(DMD/10)* (Dry Maryland Diabase rheology scaled down by 10); *B*(WOl×10)* (Wet Olivine rheology scaled up by 10); ρ = density given at reference temperatures; thermal coefficient of volume expansion =3x10^{-5} /°C. Note lower crustal density change corresponding to the 'basalt-eclogite' metamorphic phase transition. Model domain is 2000 x 600 km and comprises the lithosphere, thickness 100 km and sublithospheric mantle. Lithosphere converges asymmetrically from left at 5 cm/y. Boundary conditions on sublithospheric mantle are free slip with no material flux across the base. The sides have a small uniform outward material flux that balances the flux of lithosphere into the model. White region at left is the initial narrow

weak zone. There are no surface processes. Bold frame shows area displayed in Figures 17 and 18; note that the position of this frame migrates with time in these figures.

Fig.17. Model LHO-LS1 results showing evolution of upper mantle-scale deformation. Panels show the model materials (see Fig. 16) (dark grey regions are eclogite facies lower crust), a sparse version of the Lagrangian tracking grid, the velocity field (arrows, scale at bottom), and selected isotherms (V = H). *t* = elapsed model time, Δx = total convergence. No surface processes. Note the progressive 800 km movement of the panel windows toward the left as the model evolves, designed to keep the subducted slabs near the centre of each panel. Crustal channel flow is well developed by 30 My.

Fig.18. Model LHO-LS2 results showing evolution of upper mantle-scale deformation. Panels show the model materials (see Fig. 16) (dark grey regions are eclogite facies lower crust), a sparse version of the Lagrangian tracking grid, the velocity field (arrows, scale at bottom), and selected isotherms (V = H). *t* = elapsed model time, Δx = total convergence. No surface processes. In this case the 400 km movement of the panel windows between a) and b), and b) and c) is to the right. Crustal channel flow is restricted to the retro-crust but is well developed by 30 My.

Fig.19. Interpretation of flow modes discussed in this work in regard to the positions where they may operate in T-M space.

Fig.20. Summary diagram of the three crustal flow modes investigated in this work, together with their characteristics and the requirements for each of them to operate.

Appendix 1. Design of crustal- and upper-mantle-scale models

Philosophy of numerical approach to problem solution and model parameterisation

What are the criteria for the development of geodynamical models and how complex should they be? In the design of numerical models there is a trade-off between those that are overly simplified/specified and therefore fail to demonstrate important types of behaviour because relevant physical processes are omitted/suppressed, and those that are overly complex, displaying characteristics that are difficult to interpret owing to the large number of possible interactions. Our motivation is to illuminate the most basic physics behind orogenic evolution. We therefore choose a numerical methodology that is robust and includes the ability to solve the underlying coupled mechanical and thermal problems that operate at orogen scales. We specifically avoid 'simulation', in which the models may be over-constrained with the intent of reproducing or 'mimicking' a particular natural setting in detail. We prefer to view our models as numerical experiments designed to investigate the types of processes that occur within models of collisional orogenesis with boundary conditions that are deliberately simplified by comparison with nature. For example, the velocity boundary conditions for the crustal scale (CS) models HT1 and HT111 described in Jamieson *et al.* (this volume) and their upper mantle scale (UMS) equivalents are purposely restricted to be uniform approximations of the natural Himalayan-Tibetan system. Our thesis is that this simplified approach will reveal the underlying first-order processes.

Advantages and limitations of the CS and UMS model designs

We choose the simplest model design that is compatible with the first-order processes and features of natural orogenic systems - in this case, large, hot, collisional orogens. The CS and UMS model designs described in this paper have both limitations and advantages. Limitations include: 1) the 2D plane-strain restriction (no flow of material out of the plane of the model); 2) the crustal scale of the CS model (no mantle dynamics); and 3) the choice of basal kinematic boundary conditions. The advantages include: 1) the fully dynamical solution of the flow calculation within the CS crustal section subject to the boundary and surface process conditions, and the dynamical solution at the upper mantle scale in USM models; 2) the ability to include pressure-dependent plastic (Drucker-Prager) rheologies, corresponding to Coulomb failure and Byerlee's law, and a first-order approximation of the effects of pore-fluid pressures (CS and UMS) and a parametric model for strain softening in UMS; 3) the inclusion of thermally activated power-law viscous creep; 4) the coupled thermal-mechanical nature of the calculation; and 5) the Arbitrary Lagrangian Eulerian (ALE) formulation of the finite element problem, which both allows for sufficiently accurate calculations at medium scales within the problem domain, and includes the calculation of the evolving shape of the model domain such that orogen geometry, topography, plateau growth, surface processes and the gravitational feedback effects of changing geometry, and large deformation, are easily and naturally incorporated in the calculation.

The basic design of the ALE numerical model has been described elsewhere (Fullsack 1995; Jamieson *et al.* 2002; Beaumont *et al.* 2004) and was summarized in the Numerical Calculation section above. The same CS numerical model is used in the calculations described by Jamieson *et al*. (this volume).

Model complexity and selection and tuning of model properties

Even with the simplifications described above, the models may appear to include a large number of parameters whose values are poorly known. These can, however, be grouped into only four property sets: 1) the mechanical properties required to specify a three-layer crust (CS) and lithosphere and mantle (UMS); 2) the associated thermal properties; 3) the velocity boundary conditions; and 4) the properties of the surface processes model. All of these play important roles in natural systems and cannot be neglected in the models. Although we show only a selection of the results, they are based on extensive sensitivity analyses in which a reference model is established and then tested for its sensitivity to variations in one or, at most, two of the properties at a time - e.g., time variations in the intensity of the surface processes *f(t)*, or the spatial variation of kinematic boundary conditions.

There are three important steps in the model design: 1) the selection of a reference model; 2) the choice of parameter variations to be used in the sensitivity analysis; and 3) the assessment of the results for robust outcomes. The approach is reductionist in that a direct cause-and-effect relationship between parameter variation and model behaviour is sought. Although some of these relationships can be interpreted to be robust, the behaviour is commonly a system response involving the dynamics of one or more feedback loops that cannot be demonstrated to be uniquely related to a single input parameter.

Our experience with sensitivity analyses yields some confidence in attributing cause-and-effect relationships. It also indicates when the model outcomes become very sensitive to small variations in several input variables. In such cases, it is important to establish the range of expected variability in the model context. Equivalent natural systems can also be expected to show a range of behaviours owing to their inherent natural variability. However, it will likely be impossible to attribute a specific cause-and-effect relationship for most specific natural examples, because we commonly do not know the system properties and their associated variations accurately. Below we describe tests for some boundary conditions, the rationale for some specific parameter choices, and explain how the HT model series was developed from a simpler reference model.

Testing the basal boundary conditions in CS models

In CS models the basal velocity boundary conditions are specified kinematically to correspond to assumed behaviours of the mantle lithosphere, for example, subduction, advancing subduction, or pure shear thickening. The UMS model experiments provide an opportunity to test these assumptions by removing the specified velocities at the base of the crust and, instead, model the dynamics of the interaction between the lithosphere and underlying mantle. The observed model behaviours range from advancing double subduction, through subduction, to subduction zone retreat, and include shortening and thickening of the mantle lithosphere and various forms of convective instabilities of the mantle lithosphere (e.g., dripping, slab breakoff; Pysklywec *et al.* 2000). Mantle subduction is the preferred mode when the early stages of deformation correspond to underthrusting of one mantle lithosphere beneath the other. In the models, subduction is facilitated by a weak shear zone between the two converging lithospheres; in nature, this might be inherited from a phase of oceanic subduction prior to continent-continent collision. Distributed lithospheric contraction and thickening occurs in the absence of significant zones of weakness that could act to break the symmetry of pure shear thickening.

It cannot be demonstrated that mantle subduction necessarily accompanies continent-continent collision. However, as described in this paper, results from UMS models which predict dynamic mantle subduction are compatible with those from Himalayan-style CS models with kinematic subduction. Many UMS models, with a range of properties, exhibit subduction with combinations of subduction zone advance and retreat that are controlled by the density contrast between the mantle lithosphere and sublithospheric mantle. When the density contrast is large there is also a tendency for repeated slab breakoff events. Therefore, the possibility of punctuated subduction of mantle lithosphere must be considered, possibly associated with reversals in subduction polarity (Pysklywec 2001).

Scaling of laboratory power–law creep flow laws

We choose to base the flow laws in the models on a reference set of well constrained laboratory experimental results: wet quartz *(WQ)* (Gleason & Tullis 1995, melt-absent Black Hills quartzite), dry diabase *(DMD)* (Mackwell *et al.*1998, dry Maryland diabase), and wet olivine *(WO)* (Chopra & Patterson 1984, wet Aheim dunite; Karato & Wu.1993). Laboratory-derived flow laws are subject to significant uncertainties associated with the measurements on individual samples, the variability of measured results among samples of similar rock types, the range of deformation mechanisms, the effects of water fugacity, and the known and unknown errors in extrapolating the laboratory results to natural conditions. We have therefore chosen to limit the complexity and to base our model rheologies on a few reliable datasets in order to minimize the number of sources of error while allowing some variation in the model viscous flow properties.

Flow laws for rocks that are stronger/weaker than the base set are constructed by linearly scaling up/down the values of B^* (Eq. 3). This approach is used to approximate other material rheologies. The scaled viscosities can either be interpreted in terms of the effects of composition or the consequences of water saturated *vs*. water-poor (wet *vs*.dry) conditions. This is valid if the exponent of the water fugacity term is close to unity, and therefore the effect of water scales linearly with A in the flow law (e.g. Hirth *et al.* 2001).

Alternatively they can be interpreted as synthetic model rheologies. Given that relative ductile flow of different materials in the models is mainly a consequence of their viscosity contrast, the simple scaling guarantees that the viscosity contrast is always given by the scaling factor under the same ambient conditions. This approach simplifies the interpretation of the model results and is the principal reason for choosing it - instead of having results in which all of the parameters in the power–law creep flow law vary (Eq. 2), only the effective viscosity varies as *B** is scaled. We believe that this scaling is an appropriate way to test the sensitivity of the models to the effect of wet *vs.* dry conditions or to a moderate change in composition. For example, *B*(DRY)* is in the range 10-50×*B*(WET*), and *B*(WQx5)* approximates conditions when flow is influenced by a mineral such as feldspar that has a higher effective viscosity than wet quartz for the ambient conditions.

We choose dry Maryland diabase (*B*(DMD)*) to represent the strongest lower crustal rheology, knowing that a comparison demonstrates that *B*(DMD/10)* corresponds closely to the power-law flow properties of intermediate composition granulite (Pikwitonei granulite, Wilks & Carter 1990; Mackwell *et al.* 1998). Given uncertainties in the composition and other properties of the lower crust, we argue that a reasonable approximation of power-law creep of the lower crust can be based on proxy materials ranging from *B*(DMD)* (the strongest end member) to *B*(DMD/20)* (weak lower crust). We in no way imply that the lower crust is diabase.

Similarly, in the UMS models, ductile flow of the mantle is based on the power-law rheology of olivine-controlled rocks; we use B**(WO)* (wet Aheim dunite, Chopra & Patterson 1984) as the reference rheology, knowing that this flow law corresponds closely to that of wet olivine (Karato & Wu 1993). To a first approximation, dry olivine has an effective viscosity that is as much as 50x that of wet olivine for mantle lithosphere conditions. We therefore use *B*(WO*) for sublithospheric mantle, assumed to be water-saturated, and *B*(WOx10)* for continental mantle lithosphere that is considered to be relatively water-poor. The effect of the activation volume is not included in the calculation of the power-law creep flow laws. In the lithosphere, pressure is sufficiently low that the activation volume effect on viscosity is not significant. In the upper mantle the effect could be large, but prediction of the effective viscosity for wet olivine is complicated by the pressure and temperature dependence of water fugacity and whether the system behaviour is open or closed (Karato & Jung 2003). For the purposes of the demonstration models we omit both of these effects, but limit the sublithospheric minimum viscosity to 10^{19} Pa.s, which is somewhat larger than the predicted water-saturated values (Karato & Jung 2003).

Design of Himalayan-Tibetan (HT) models

The HT series models were developed from a large-hot-orogen CS reference model similar to Model 1 of Beaumont *et al.* (2004) but with no melt weakening or erosion. The reference model has V_P = 2.0 cm/y and $V_S = V_R = 0$. The undeformed crust has ϕ_{eff} = 15° throughout, and comprises a 25 km thick upper/mid-crustal layer with *B*(WQ)* and a 10 km thick lower crust with *B*(DMD)*. The lower crust is not subducted, there is no melt weakening, and surface processes are not included. Thermal properties are those used by Jamieson *et al* (2002) and Jamieson *et al.* (2004b). The models contains two layers with contrasting heat production, A_1 = 2.0 μW/m^3 (0-20 km) and A_2 = 0.75 μW/m^3 (20-35 km). These values were chosen to represent continental margin crust (Jamieson *et al.* 2002); upper crustal heat production, in particular, lies within the range reported from GHS lithologies (e.g. Huerta *et al.* 1998, and references therein). As described in the Numerical Calculation section above, values of other thermal parameters (K, κ, C_P, ρ) are identical in both layers and lie in the mid-range of those normally attributed to continental crust (e.g., England & Thompson 1984).

Results from the reference model show significant departures from the first-order properties of the Himalayan-Tibetan orogen. A number of physically justifiable modifications were therefore made which led to the HT series of models, from which representative model HT1 was subjected to detailed analysis (Beaumont *et al*. 2004; Jamieson *et al.* 2004b, this volume). The five essential modifications incorporated into the HT series models in order to produce model results compatible with observations are listed below. Model thermal properties were not adjusted.

1) *Change velocity boundary conditions.* For consistency with estimates of average India-Asia convergence velocity, the HT series models have V_P = 5 cm/y. The models are viewed in the fixed Asia reference frame, $V_R = 0$, with advancing subduction, V_S = 2.5

cm/y. Royden *et al.* (1997) and Beaumont *et al.* (2004) demonstrated that advancing subduction is required to reproduce the general planform geometry of the Himalayan-Tibetan system and the surface position of the Indus-Tsangpo suture. For reasons noted above, velocity boundary conditions remain constant during each of the CS model experiments.

2) *Subduct lower pro-crust*. Accumulation of lower pro-crust in the model orogen produces a large lower crustal antiform, inconsistent with data from the Himalayan-Tibetan orogen (e.g., Model 1 of Beaumont *et al.* 2004). The lower crustal layer on the pro-side of the model system (corresponding to India) is therefore subducted along with the pro-mantle lithosphere. This is consistent with mechanical coupling between strong lowermost crust and upper mantle in mature continental crust, and with lithosphere-scale interpretations of seismic data from the orogen (Owens & Zandt 1997). Because the lower pro-crust is detached and subducted at the S-point, it behaves like the mantle directly beneath it and is not deformed during model evolution. As the overlying crust thickens and heats up, it becomes mechanically decoupled and detached from the lower crust, which is overridden as the orogen propagates towards its foreland.

3) *Include melt weakening.* As shown by Beaumont *et al*. 2004 (Model 3 *vs* Model 1), models without melt weakening produce, at best, inefficient channel flows restricted to the region beneath the central model plateau. Including a parameterised viscosity reduction over the temperature interval associated with melting (Beaumont *et al.* 2001, 2004, Jamieson *et al.* 2002) produces efficient channel flows extending to the plateau edge. This is consistent with seismic evidence that some melt is present under the present-day Tibetan plateau (Nelson *et al*. 1996; Klemperer this volume) and with observations that GHS gneisses (exhumed equivalents of postulated channel) are typically migmatitic. In HT models, melt weakening is restricted to the middle and upper crustal (quartzo-feldspathic) layers and does not affect the lowermost (granulitic) crust.

4) *Include surface denudation*. In the absence of erosion, the channel flow zone "tunnels" into the surrounding crust at a rate controlled by its thickness and the temperature field (Royden 1996; Beaumont *et al*. 2004, Medvedev & Beaumont this volume). In order to exhume the channel it is necessary to erode the plateau flank. In the HT models, surface denudation is controlled by the interaction of surface slope, a spatial function (*g(x)*), and

a time function (*f(t)*). Local surface slope is calculated within the model. To a first approximation *g(x)* is a measure of the spatial variation of aridity (0 = dry, 1 = wet) across the model, and *f(t)* combines the effects of long term climate variations, the bedrock incision rate constant, and a parameter that scales the model surface slopes, which are determined on a 10 km spatial resolution, to include higher riverbed slopes at smaller scales. A more detailed denudation model is not justified because the model is cross-sectional, and therefore cannot represent planform drainage patterns, and the scaling effect in *f(t)* for local slopes at less than 10 km spatial resolution is not known accurately.

All HT models are run for an initial set-up phase (0-24 My; 54-30 Ma) without surface denudation. This is not a significant factor in the later stages of model evolution (the focus of our work to date), and is designed to achieve a model state with an embryonic plateau and mid-crustal channel flow as a precursor to testing model sensitivity to denudation. The results are similar with moderate denudation during the set-up phase but the times to develop the plateau and channel flow are somewhat longer. In model HT1, erosion rate is high from 24-39 My (30-15 Ma), which initiates efficient channel extrusion, and then declines gradually from 39-54 My (15-0 Ma) towards present-day values. Similar model results are obtained using somewhat different denudation functions (e.g. Model 3 of Beaumont *et al.* 2001). However, successful models require a period of rapid erosion (*f(t)* large) after 24 My (30 Ma) in order to initiate channel exhumation, and a decline from the maximum rate (*f(t)* decreasing) in the last 15-20 My of model evolution in order to produce model ages for peak metamorphism and cooling that lie within the observed range. As noted by Jamieson *et al.* (2004), "GHS" cooling ages predicted by HT1 are too young, suggesting that recent erosion rates should be even lower. With the additional provenance and detrital mineral data that have recently become available (e.g. DeCelles *et al.* 2004; Amidon *et al.* 2005; Najman *et al.* 2005; Najman 2006), different denudation functions might be chosen for a future series of models. The original HT1 design is retained by Jamieson *et al*. (this volume) in order to complete the analysis of that particular model.

5) *Include 3 crustal layers*. The modifications noted above lead to the development of a Himalayan-scale model orogen with extrusion of a mid-crustal channel on timescales of

50-55 My. However, a crustal structure comprising three laterally continuous layers with contrasting mechanical properties produces significant improvements in the model. In particular, a weak upper crustal layer that is capable of detaching from underlying middle crust allows the formation of an asymmetric overthrust structure at the orogenic front and domes in the region between the plateau flank and the suture (Beaumont *et al.* 2004; Jamieson *et al.* this volume). The rheology of the uppermost layer (0-10 km) is given by *B*(WQ)* with ϕ_{eff} = 5°, representing sedimentary rocks of the upper crust with high pore fluid pressures. The middle crustal layer (10-25 km) uses *B*(WQx5)* with ϕ_{eff} = 15°, representing quartzo-feldspathic granitic and/or metasedimentary rocks. As described above, the upper and middle crustal layers are subject to melt weakening where $T \geq$ 700°C. The rheology of the lower crust (25-35 km) is given by *B*(DMD)*, with ϕ_{eff} = 15°, representing lower crustal granulite. Similar results are obtained with *B*(DMD/5).* The lower crustal layer is not subject to melt weakening. Some HT series models use variations on this simple 3-layer structure, which are described where specific model results are presented.

Essential model requirements for channel flow

We use the terms channel flow and extrusion to describe the general process of orogen-scale, confined, pressure-driven flow (analogous to pipe flow, Turcotte & Schubert 1982, p.237) and the ejection of the channel material toward the surface near the end of the flow zone. In order to generate channel flow in the model, the only requirements are reduced viscosity, $\eta_{eff} \leq 10^{19}$ Pa s, and a pressure differential sufficient to drive a flow with that viscosity. In the HT models, both the pressure differential and the reduced viscosity result from crustal thickening. The pressure differential comes from the potential energy difference resulting from the contrast in crustal thickness and elevation between the plateau and the foreland, and the viscosity reduction is associated with high temperatures generated by heat production in thickened crust. Beneath the plateau, material flux through the channel is related to its thickness and viscosity (scales with h^3/η; e.g. Royden 1996) and the rate of channel propagation is limited by the rate at which adjacent material becomes hot and weak enough to be incorporated into the advancing channel flow (Beaumont *et al*. 2004; Medvedev & Beaumont this volume).

The active or previously active (fossil) channel is exhumed by focused surface denudation. Extrusion between coeval thrust- and normal-sense shear zones occurs where material in an active channel ($T \geq 700°C$) is pumped or forced towards the surface - by analogy with pipe flows, the surface represents the open end of the pipe. Since the temperature at the model surface is 0°C, channel material cools during extrusion at a rate determined largely by the rate of denudation. In the models, and probably also in nature, the geometry of the channel is significantly modified during extrusion. In the model, deformation superimposed on the channel material at this stage generally involves flattening and thinning. By implication, structures in natural exhumed channels should record features formed during active channel flow as well as features superimposed during extrusion, exhumation, and cooling. It might therefore be difficult to determine unambiguously whether or not channel flow has occurred from structural analysis of specific exhumed sections.

Model-data comparisons

What are the most effective tests of the models? The feasibility of any numerical model for orogenesis must be tested against data from real orogenic systems. Conversely, the feasibility of conceptual models based on geological or geophysical data, and of kinematic models based on predefined geometries, should be tested against the physics of the system as a whole. Are the assumptions physically realistic? In either case, the tests should be designed to reflect the first-order properties of the model system on the appropriate scale. If the models fail the first-order tests, second-order features are irrelevant. If the models pass the first-order tests, it must be determined whether second-order model predictions are robust, and therefore testable, and whether the second-order features themselves are well enough characterised to constitute valid tests of a specific model.

Given that orogenesis is a response to the behaviour of the lithosphere during convergence, the present models are designed on the scale of the crust and upper mantle. This imposes numerical limitations on model resolution and there is a corresponding limit to the scale at which model predictions can be reliably compared with data from specific orogenic transects. A further limitation on model-data comparisons is the 2D, plane-strain, nature of the numerical models presented here. The Himalayan-Tibetan system displays remarkable along-strike continuity (e.g., Hodges 2000), which allows reasonable first-order model-data comparisons for the central part of the orogen. However, where 3D effects are important, e.g., in the vicinity of the Himalayan syntaxes, specific model-data comparisons become tenuous.

In comparing our model results with data from natural orogens, we first assess consistency with crustal or lithospheric-scale features before making comparisons with specific seismic, structural, metamorphic, stratigraphic, or geochronological datasets. In compiling geological or geophysical data, we look for regional-scale consistency in order to distinguish general (first-order) properties of the system from those controlled by local features. Similarly, matching the details of a particular type of data (e.g. a specific P-T-t path) is less important than consistency with combinations of data (e.g. P-T-t path style combined with peak grade profiles and geochronology).

The first-order test of the channel flow model is the existence of mid-crustal channels with large-scale flows characterised by velocities on the order of 1 cm/y. This has not yet been detected directly. In the Himalayan-Tibetan system, indirect evidence for channel flow includes a variety of geophysical data from the Tibetan plateau, as summarised by Klemperer (this volume), the magnetotelluric evidence (e.g. Unsworth *et al.*, 2005) and a range of geological data summarised by Jamieson *et al.* (2004b). While indirect evidence may not constitute a diagnostic test, the ability of the homogeneous channel flow model to account for a wide array of disparate features of the orogen suggests that the simple model captures many essential elements of the behaviour of the system. We conclude that channel flow models in general provide a reasonable first-order explanation for the thermal-tectonic and lithological evolution of the Himalaya and southern Tibet.

In exposed mid-crustal levels of ancient orogens, a number of geological observations could constitute tests for the former existence of channel flows. These include: 1) the presence of coeval normal- and thrust-sense shear zones bounding a regionally extensive zone of migmatite or some other material inferred to have had low viscosity (relative to underlying and overlying rocks) at the time the shear zones were active; 2) a transition from an "inverted" metamorphic sequence across the basal thrust-sense shear zone into a "normal" metamorphic sequence across the upper normal-sense shear zone; 3) evidence that ductile flow in the low viscosity zone post-dated compressional deformation in overlying crust by ca. 20-25 My (time needed to initiate melt-weakening in thickened crust); 4) discontinuity between upper and lower crustal structures across the ductile flow zone; 5) evidence for substantial lateral transport of low-viscosity material along structures that were shallow-dipping at the time that they formed.

Beaumont *et al.* (2001, 2004, this paper) and Jamieson *et al.* (2004b, this volume) have demonstrated that both CS and UMS models are sensitive to small variations in parameters such as crustal strength, denudation history, and upper mantle density. Within the range of natural variability of these parameters, the model system can respond in different ways to produce a variety of features observed in different places and/or times in the evolution of the orogen. The resulting variability does not extend to its first-order features - i.e. the generation and exhumation of mid-crustal channel flows - but can produce significant differences in the surface expression of the underlying processes. Under these circumstances a model could potentially be "tuned" to achieve a desired effect, for example to explain the details of a specific transect. As discussed above, model tuning to fit second-order features provides little or no insight into processes, and the resulting match does not constitute a valid test of the model.

However, far from being a weakness of the HT model series, its sensitivity to variations in parameters that are demonstrably variable in nature should be regarded as one of its strengths. This is in itself an important test of the model. Models that fail to predict natural variability are inadequate. It follows that the expectation that one specific model should explain all the features of an orogen is wrong, and conversely there is no unique model against which all observations should be compared.

Table 1. Parameters used in models (see also Figure 2).

Parameter	Meaning	Value(s)
Parameters and nominal values		
a) Mechanical parameters		
ρ_{crust}	crustal density	2700 kg/m^3
ρ_{mantle}	mantle density	3300 kg/m^3
D	flexural rigidity (isostasy model)	10^{22} Nm
	crustal thickness	35 km
	lower crustal thickness	see below
θ	subduction dip angle	20^o
ϕ_{eff} (0 - 10 km)	effective internal angle of friction	5°
ϕ_{eff} (10 - 35 km)		15^o
C	cohesion	10 MPa
P	solid pressure	Pa
$J_2^{/}$	second invariant of the deviatoric stress tensor	Pa^2
$\eta_{eff}^{v} = B^* \cdot (\dot{I}_2^{/})^{(1-n)/2n} \cdot \exp[Q/nRT_K]$	general equation for effective viscosity	
$\dot{I}_2^{/}$	second invariant of strain rate tensor	s^{-2}
R	gas constant	8.314 J/mol^{o}K
T_K	absolute temperature	^{o}K
B, n, Q* as below		
WQ (0 – 10 km)	wet Black Hills quartzite flow law [after *Gleason and Tullis*, 1995]	n = 4.0 B *= 2.92 x 10^6 Pa.s$^{1/4}$ Q = 223 kJ/mol
WQ x *5* 10 – 25 km or 10 – 20 km (see below)	modified wet Black Hills quartzite flow law	*B*= B* (WQ) x 5* (etc.)
DMD	dry Maryland diabase flow law [after *Mackwell et al.,* 1998]	n = 4.7 *B**= 1.91 x 10^5 Pa.s$^{1/4.7}$ Q = 485 kJ/mol
DMD/f (see below)	scaled dry Maryland diabase flow law	*B* = B* (DMD) / f*
'melt weakening'	linear reduction in effective viscosity over T range 700-750°C for WQ only	η_{700} = flow law value η_{750} = 10^{19} Pa.s
	length of Eulerian model domain	2000 km
b) Crustal scale models basal velocity boundary conditions		
V_P	pro-side (convergence) velocity	1.5 cm/y
V_R	retro-side velocity	-1.5 cm/y
V_S	S-point velocity (subduction advance)	0 cm/y

c) Thermal parameters

C_P	heat capacity	750 $m^2/^oKs^2$
K	thermal conductivity	2.00 W/mºK
κ	thermal diffusivity ($\kappa = K / \rho Cp$, where $\rho Cp = 2 \times 10^6$)	1.0 x 10^{-6} m^2/s
T_s	surface temperature	0ºC
T_a	temperature at lithosphere/ asthenosphere boundary	1350ºC
q_m	basal mantle heat flux	20 mW/m^2
q_s	initial surface heat flux	71.25 mW/m^2
A_1 (0-20 km)	upper crustal heat production	2.0 $\mu W/m^3$
A_2 (20-35 km)	lower crustal heat production	0.75 $\mu W/m^3$

d) Crustal scale models surface denudation

slope x $f(t)$ x $g(x)$	denudation rate (m/y)	
slope	local surface slope measured from finite element mesh	
$f(t)$	time function specifies how denudation rate (m/y) varies with time when $g(x)$ and slope = 1	constant
$g(x)$	spatial function specifies how denudation rate varies with position x	$g(x)$ = 0 = arid $g(x)$ = 1 = wet

e) Specific model parameters – Crustal Scale Models

LHO-1
Lower crust (25 – 35 km) *B* (DMD/5)*
15°

LHO-2
Lower crust (25-35 km)
Alternating 250 km long blocks of *B* (DMD)*
B (DMD/10)*

LHO-3
Lower crust (25-35 km)
250 km long blocks arranged symmetrically with respect to S. Blocks have properties
B (DMD), B* (DMD/4), B* (DMD/8), B* (DMD/12), B*(DMD/16), B* (DMD/20)*
Order is from external to internal part of model.

Table 2. Parameters used in Upper Mantle Scale (UMS) models (where different from those of the Crustal Scale (CS) models).

(see also Figure 16) LHO-LS1 and LHO-LS2

Parameter	**Meaning**	**Value(s)**
Parameters and nominal values		
a) Geometry		
	Model domain	2000 x 600 km
	Eulerian mesh	101 x 201
b) Mechanical parameters		
ρ_{uc}	upper crust density at 194°C*	2800 kg/m^3
ρ_{lc}	lower crust density at 457°C*	2950 kg/m^3
ρ_e	nominal lower crust density when transformed to eclogite facies	3100 kg/m^3
ρ_{ml}	mantle lithosphere density at 937°C*	
	in model LHO-LS1	3300 kg/m^3
	in model LHO-LS2	3310 kg/m^3
ρ_{um}	uniform sublithospheric upper mantle density	3260 kg/m^3
* Initial average temperature of this model layer		
ϕ_{eff} (0 - 28 km)	effective internal angle of friction (strain softens linearly over range 0.5 → 1.5 of second invariant of strain)	15° → 2°
ϕ_{eff} (28 - 34 km)	effective internal angle of friction (strain softens linearly over range 0.5 → 1.5 of second invariant of strain)	15° → 2°
ϕ_{eff} (34 - 600 km)	effective internal angle of friction (strain softens linearly over range 0.5 → 1.5 of second invariant of strain)	15° → 2°
C_{uc}	cohesion	10 MPa
	viscous flow laws (see Table 1)	
WQ x *5* (0 – 28 km)		$B^* = B^*$ (*WQ* x *5*)
DMD/10 (28 – 34 km)		$B^* = B^*$ (*DMD/10*)
WO x *10* (34 – 100 km)	scaled olivine flow law	$B^* = B^*$ (*WO* x *10*)
WO (100 – 600 km)	olivine flow law	$B^* = B^*$ (*WO*)
WO	wet Åheim dunite (olivine) flow law [after *Chopra and Paterson*, 1984]	$n = 4.48$ $B^* = 7.75 \times 10^4$ Pa.s$^{1/4.48}$ $Q = 498$ kJ/mol $V^* = 0$
	minimum effective viscosity in sublithospheric mantle	10^{19} Pa.s

c) Upper mantle scale models velocity boundary conditions

Velocity boundary conditions		
V_P (0 – 100 km)		5 cm/y
V_R (0 – 100 km)		0 cm/y
	small flux through side boundaries (see text)	
	other boundaries, free slip; upper surface, free surface	

d) Thermal properties

	melt weakening (see Table 1 and Figure 2)	
K	thermal conductivity	2.00 W/mºK
κ	thermal diffusivity	8.0 x 10^{-7} m^2/s
κ_{um}	thermal diffusivity of sublithospheric upper mantle (adiabatic temperature gradient)	3.2 x 10^{-5} m^2/s
T_s	surface temperature	0ºC
T_a	temperature at lithosphere/ asthenosphere boundary	1350ºC
T_b	initial temperature at model base	1493ºC
q_m	basal mantle heat flux	20 mW/m^2
q_s	initial surface heat flux	70.5 mW/m^2
α_V	volume coefficient of thermal expansion	3 x 10^{-5}/ºC
A_1 (0-20 km)	upper crustal heat production	2.0 $\mu W/m^3$
A_2 (20-34 km)	lower crustal heat production	0.75 $\mu W/m^3$