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 , mhnguyen@dal.ca,*

[2] *Department of Earth Sciences, Dalhousie University, Halifax, Nova Scotia, Canada B3H 3J5, beckyj@dal.ca*

[3] *Institute for Geological and Nuclear Sciences, Lower Hutt, New Zealand, s.ellis@gns.cri.nz*

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 end 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.

(223 words needs to be 200 or less)

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); and 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, 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*. 2004).

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 surface conditions and provides insight into the range of stellar evolution. The intent of the T-M diagram (Fig.1) is 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 crustal lithosphere, and the excess heat content or temperature of the orogen relative to the same undeformed crust/lithosphere with standard heat production. The T-M diagram provides a first-order classification of orogen types (e.g. Dwarfs, Giants etc., Fig 1a) and insight into the processes that occur in these orogens. In addition, the evolution of orogens, like stars, can be represented by evolutionary paths in the T-M diagram. An ‘orogenic main sequence’ (MS) (Fig.1) runs from bottom left to top right and describes orogens that have excess conductive steady-state, temperatures complementing their excess crustal thickness. 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. This means that 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 T=M line. The convex up curvature for large orogens expresses the trend to a gravitational limit on the maximum crustal /lithospheric thickness as it becomes hot and weak. 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 New Zealand, Pyrenees and Alps plot in the lower left part of diagram (Fig.1b), and 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), 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 mid- and/or lower crust. Notable exceptions are volcanic arcs, which may be small, but sufficiently hot and weak to undergo crustal flows of the type we describe below under 'large-hot' orogens.

Large-hot orogens plot at the other end of the orogenic main sequence and are both massive and hot (Giants and Super Giants, Fig.1a), 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 the eastward growth of the Tibetan plateau, as the channel tunnels outward (Clark & Royden 2000, and refs therein), and the ductile extrusion of the Greater Himalayan sequence (Grujic *et al*. 1996, 2002; Beaumont *et al*. 2001, 2004; Jamieson *et al*. 2004). 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), can 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 have been described in the earlier papers referenced above and are shown on Figure1c. 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 has Heterogeneous Channel Flow, which incorporates lower crustal blocks within the channel; Mode 3, entitled 'Hot Fold Nappes', measures the response of the model orogen to the insertion of progressively stronger blocks of lower crust. We interpret the three modes as end members of a continuum of gravitationally and tectonically driven 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 show channel flows and address the related question concerning the fate of lithospheric mantle during continent-continent collision.

Numerical Calculations of Crustal Flows in Large-Hot Orogens

The explanation of the numerical calculations follows Beaumont *et al*. (2004) and is included here for completeness. We model the development of large-hot orogens using a two-dimensional (2D) finite element code that assumes plane-strain conditions in a cross-section through the orogen. The code computes both thermal and mechanical evolution subject to velocity boundary conditions applied at the sides and base of the modeled region. Thermal-mechanical coupling occurs through the thermal activation of viscous power-law creep in the mid and lower crust. The model properties are similar to those described by Beaumont *et al*. (2004) and Jamieson *et al*. (2004).

The model 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). The associated temperature field is calculated for the whole model domain. Model parameters and values are given in Table 1. In the models

described here, both mantle lithospheres, termed 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 the symmetric convergence were chosen to give the most generic 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 with constant velocity as a slab with constant dip, θ (Fig. 2). The model can also be interpreted in an alternative reference frame (Beaumont *et al*. 2004) 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, $V_R = 0$.
The mechanical model used to calculate the velocity field and deformation (Fullsack 1995) uses 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 velocity boundary conditions described above (Fig. 2a).

The corresponding thermal evolution includes diffusion, advection, and radioactive production of heat and is calculated on the same Eulerian finite element mesh by solving the heat balance equation. The advection velocities are prescribed kinematically in the mantle and calculated dynamically in the crust as described above. A flexural isostatic compensation resulting from changes in crustal thickness is calculated from the elastic flexure of a beam embedded in the model at the base of the crust.

The surface processes model specifies the local erosion rate as $\dot{e}\,(t,x)$ = slope $\times$ $f(t)$ $\times$ $g(x)$ 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 that varies spatially (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 models described here $f(t)$ is constant.

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 $(J_2^{/})^{1/2} = P\sin\phi_{eff} + C\cos\phi_{eff}$ (see Table 1 for symbols) where the value of ϕ_{eff} is defined to include the effects of pore fluid pressures through the relation $P \sin\phi_{eff} = (P - P_f) \sin\phi$, where ϕ =30° for dry frictional sliding conditions (approximately Byerlee's law conditions) when the pore fluid pressure, $P_f = 0$. For hydrostatic fluid pressures and typical crustal densities ϕ_{eff} is approximately 15°.

The incompressible plastic flow becomes equivalent to a viscous material (Fullsack 1995; Willett 1999) such that $\eta_{eff}^{P} = (J_2^{/})^{1/2} / 2(\dot{I}_2^{/})^{1/2}$. 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 viscous stress is less than the plastic yield stress for the local ambient conditions. Under these circumstances the power law creep effective viscosity, $\eta_{eff}^{V} = B^{*} \cdot (\dot{I}_2^{/})^{(1-n)/2n} \cdot \exp[Q / nRT_K]$ (see Table 1 for symbols) and the values of B^*, n, *and* Q are based on laboratory experiments (Table 1) with A values converted to B^* assuming cylindrical creep tests. The rheology of the upper crust (0-25 km in the initial configuration) is based on the 'Wet Black Hills Quartzite' flow law (Gleason & Tullis 1995). In the model experiments (Table 1) we incorporate the effects of a mix of quartz- and feldspar-dominated viscous flow by scaling the value of *B*(WQ)* by 5 to represent crust that has somewhat higher viscosity than the standard Gleason & Tullis (G-T) flow law.

The rheology of the lower crust (25-35km initially) is based on the 'Dry Maryland Diabase' flow law (Mackwell *et al.* 1998) (Table 1). The value of *B*(DMD)* is also scaled to achieve a range of effective lower-crustal strengths. The rheological structure can be considered to represent a three-layer crust comprising two layers of quartzo-feldspathic rocks, the upper one with weak frictional-plastic properties, $\phi_{eff} = 5°$, and the lower one with standard frictional-plastic properties, $\phi_{eff} = 15°$, underlain by an intermediate dry granulitic lower crust. There is no strain softening in the models described in the next section.

The most important additional model property is an extra increment of viscous weakening in the upper and middle crustal material 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 $T>750°C$. This weakening approximates the reduction in the bulk viscosity caused by a small amount of *in situ* partial melt, estimated to be approximately 7% at the melt connectivity transition (Rosenberg & Handy *in press*), and does not correspond to the much larger decrease in effective viscosity that may occur at the solid to liquid transition. The 'melt weakening' used in the models described here amounts to a maximum of approximately a factor of 10 decrease in effective viscosity, probably a conservative estimate for melt weakening. In the models lower crust is interpreted to be refractory mafic to intermediate granulite and not prone to melting at the temperatures achieved in the models. In all instances, the model deforms according to the mechanism that produces the lowest level of the second invariant of the deviatoric stress for the prevailing conditions.

Both the mechanical and thermal calculations are carried out for each model timestep (Fig. 2). The upper 20 km of the crust has a uniform radioactive heat production, $A_1 = 2.0\ \mu W/m^3$ that is higher than the corresponding heat production, $A_2 = 0.75\ \mu W/m^3$ in the lower crust (Jamieson *et al*. 2002) For each model run, the initial steady-state temperature field is calculated at the lithospheric scale with a basal heat flux, $q_m = 20\ mW/m^2$, a surface temperature of $0°C$, and no heat flux through horizontal side boundaries. The lithosphere-asthenosphere boundary is defined to be at the $1350°C$ isotherm. For these conditions and thermal conductivity, $K = 2.00\ Wm^{-1}\ C^{-1}$, the initial surface heat flux $q_s = 71.25 mW/m^2$, and the Moho temperature is 704°C. The effect of a precursor phase of oceanic subduction, which is 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).

Model Results

In this section we describe results from three models. These models are LHO-1, which illustrates Mode 1, Homogeneous Channel Flow; LHO-2 illustrating Mode 2, Heterogeneous Channel Flow and LHO-3 illustrating Mode 3, Hot Fold Nappes. All three models are the same except for the properties of the lower crust, which are 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 composed of Dry Maryland Diabase scaled down by a factor of 5. This scaling achieves an effective strength that is intermediate between the very strong DMD and typical intermediate granulites (e.g. Pikwitoniei granulite, Mackwell *et al*. 1998, which has an effective strength with *B*(DMD/10))*. The pro- and retro- sides of the model (Figs. 3 and 4, a and b) 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 velocity and temperature fields. Convergence is symmetric with $V_P = 1.5cm/yr$ and $V_R = -1.5cm/yr$.

During the initial 25My 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 crust, the development of a sub-horizontal shear zone in the mid-crust, and the viscous decoupling of the relatively weak lower mid-crust from the stronger *B*(DMD/5)* lower crust (Figs. 3 and 4, a and b, 20My). The lower crust is weakly sheared and the basal boundary condition forces it to detach and thicken near the centre of the model.

This effect is probably not realistic and Beaumont *et al*. (2004) argued that lower crust is most likely subducted during orogenesis because orogenic 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 radioactive heating. 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. 4, a and b, 20 and 30My). Thermal re-equilibration by radioactive self-heating and thermal diffusion occurs with a timescale of close to 20My during which time the temperature in the lower crust reaches 800°C (Fig. 4). This self-heating timescale is much shorter than the 50-100My required for lithospheric-scale thermal relaxation.

At approximately 25My, channel flow starts in the retro-midcrust, where temperatures are above 750°C and the mid-crust is melt weakened, and is soon followed by an equivalent flow in the pro-crust (Figs. 3 and 4 a and b, 30My). The flows develop against the thickened lower crust, but this core does not cause the flow by acting as a backstop because the same flow regimes occur in models where it is absent (Beaumont *et al.* 2004). The minor asymmetry in the flow is a consequence of the erosion on the pro-side of the model. The channel flows subsequently tunnel outward such that their tips evolve with the temperature field, coinciding with the 750°C isotherm. These positions are also close to the edges of the orogenic plateau that develops in the centre of the model (Figs. 3 and 4, b, 30-60My). The only significant difference between the two sides is the erosional uplift and exhumation of the pro-flank, which results in tectonic thickening of the mid-crust, but is not sufficiently aggressive to exhume the channel, which continues tunnelling in the mid-crust.

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 case of 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. 5 and 6, a and b) provides some insight into the effect of variations in lower crustal properties on the thermal-tectonic style of the models. Our particular focus concerns the relative styles of deformation of the mid- and lower crust and the difference by comparison with the homogeneous lower crust, LHO-1.

Model LHO-2 is similar to model LHO-1. The only difference is that the lower crust comprises alternating 250km wide zones with the standard Dry Maryland Diabase rheology, *B*(DMD),* and with the same rheology but with *B*(DMD/10)*. This scaling means that the high and low viscosity regions in the lower crust have a nominal viscosity contrast of 10, designed to correspond approximately to the difference between dry, refractory mafic lower crust and intermediate (e.g. Pikwitonei, see above) granulite lower crust. However, this factor of 10 contrast will be modulated by the nonlinear effect of power-law flow and temperature variations. The lower crustal strong blocks are therefore nominally a factor of 2 stronger than the lower crust on LHO-1 and the weaker blocks are a factor of 5 weaker.

The results for model LHO-2 show a complexly deformed crust that can be understood as the superposition of two main deformation phases. The first phase is the process that activates and deforms the zones of weaker lower crust. This deformation occurs in the transition zone between the foreland and the plateau (Fig. 5b) and has a style that 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 the neighbouring regions of strong lower crust (Fig. 5, 30-50My). Shears at the leading edges of the tongues propagate upward through the crust and the allochthonous tongues and their overburden become packages of uplifted and transported crust. Where lower crust is evacuated it is replaced by subsiding mid-crust and these regions preferentially shorten and thicken during further orogen contraction (e.g. vertical markers –2 to –3 and -4 to –5, Fig. 5b).

In the second phase 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. 5 a and b, 50-60My). 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. 5a, 30-60My) that is similar to that of LHO-1.

The implication is that heterogeneous lower crust may make the geometry and composition of the channel flows similarly heterogeneous. However, widespread channel flows can also develop even under these circumstances.

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 7 and 8. 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)* through *B*(DMD/10)*, a representative granulite (Mackwell *et al*. 1998, Pikwitonei granulite) to half this value, *B*(DMD/20)*. The entire lower crust has $\phi^{P}_{eff} = 15°$, but the 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 absorbed 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. 7 a and b, 30My). 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. 7 and 8, a and b, 30My). Diachronous thickening of the radioactive crust is relatively fast and creates thermal disequilibrium owing to the vertical stretching of the thermal regime (Fig. 8a, 30My). 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. 7 and 8, b, 30-40My). Phase 2 is also diachronous and typically takes an additional 20My in these models after crustal thickening ends (Fig. 8b, 40My). Phases 1 and 2 occur sequentially in each part of the crust 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 the there is no significant change in deformation style. However, the *B*(DMD/4)* block offers some resistance to decoupling and, therefore, forces additional contraction on the interior of the system which responds by developing large-scale lower-crustal folds (Fig. 7, a and b, 40My). 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 to develop large-scale, gently inclined, ductile fold nappes rooted at the Moho (Fig. 7b, 50My); some of these are then expelled over the indentor and either inserted into the middle crust (Fig. 7b, 60My) and/or exhumed to the surface by erosion (Fig. 7a, 60My). Comparison of the pro- and retro-sides of the model demonstrates that the primary effect of surface denudation during Phase 3 is to determine the relative amount of uplift and exhumation of the fold nappes versus their horizontal transport once inserted into the mid-crust. If there is little or no erosion, the nappes remain buried and are transported together with the overlying crust, which shows little deformation associated with nappe insertion (Fig. 7b, 60My). As explained below, the Hot Fold Nappe style of crustal flow is favoured by weak lower crust in the interior of the orogen. The level of weakening is related to the incubation time, the length of Phase 2 for each part of the model crust (see Discussion below).

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 that correspond to suduction of the pro-mantle lithosphere, and kinematically prescribed variations on this theme, including subduction zone retreat and advance. The latter is also referred to as ablative subduction (Tao & O'Connell 1992; Pope & Willett 1998). However, it is important to determine whether subduction of the mantle lithosphere is dynamically consistent with the assumed properties of the model or whether other types of deformation may occur. Upper mantle scale numerical thermal-mechanical models with viscous-plastic rheologies have been presented by Pysklywec 2001; Pysklywec *et al.* 2000, 2002; Pysklywec & Beaumont 2004). These results demonstrate several modes of mantle lithosphere deformation including subduction with both advance and retreat components, double subduction, and slab breakoff. However, most of this work focused on the early stages of continental-continent collision. Here, the focus is on flow modes in large-hot orogens, therefore we report on two model experiments that illustrate the types of mantle lithosphere behaviour that may occur during prolonged continent-continent collision.

Description of Model Experiments

Both models include the lithosphere and upper mantle (Fig.9 and Table 1) and are laterally uniform except for a narrow weak zone in the crust and uppermost mantle designed to represent a simplified suture that acts to initiate underthrusting . We do not consider a precursor phase of oceanic subduction in these experiments. The models are designed to correspond approximately to the collision of India with Asia, so the boundary condition has pro-lithosphere, equivalent to India, converging from the left at a uniform velocity of V_P*=5 cm/yr* against a stationary retro-lithosphere, V_R *=0 cm/yr* corresponding to Asia. In contrast to the models described above, V_S is not specified but is determined by the dynamical evolution of the model. The sublithospheric mantle parts of the sides and base have free slip boundary conditions and 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. No surface erosion or deposition occurs in these models, which are designed for comparison with the simple tunneling mode of channel flow (Beaumont *et al.* 2004, Figs.12a and 13a). The only difference between the 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. The results show that the model behaviour is very sensitive to this 0.3% difference in density. The crust is similar to that in the crustal-scale models, except the internal angle of friction in the frictional-plastic rheology strain softens from $\phi_{eff} = 15°$ 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.

The models also include a change in density of the lower crust from 2950 to 3100 kg/m^3 when pressure and temperature conditions cross the boundary corresponding to the basalt-eclogite metamorphic transition. The density increase is relatively small because only a fraction of the crust is considered to transform to high-density eclogite.

Upper Mantle Scale Model Results: Models LHO-LS1 and LHO-LS2

During the initial stages of convergence in both models (e.g. Fig.10a for LHO-LS1) the mantle lithosphere asymmetrically underthrusts and subducts at a relatively low angle in a ‘plate-like’ manner with little internal deformation. We use the ‘pro- retro-‘ terminology because this behaviour is similar to the prescribed subduction in the crustal scale models (e.g. Beaumont *et al*. 2004). By 9My the behaviours of the subducted slabs diverge. The denser mantle lithosphere in LHO-LS2 begins to sink, in addition to subduct, and the lower part of the slab steepens and dips at high angle (Fig. 11a). In contrast the slab in LHO-LS1 decreases its dip and resists subduction and the retro-mantle lithosphere deforms to accommodate the contraction. The latter is the first stage in what develops into advancing (Fig. 10b) and then double subduction (Fig.10c) during which the subduction point advances dynamically such that there is net subduction zone advance of approximately 700km between 9 and 33My, corresponding to an average V_S = 3.2 cm/y. At approximately 30My, the buoyancy of the double slab becomes sufficiently negative that viscous necking starts, leading to breakoff of the double slab at 42My, (Fig.10d) by which time the subduction point has advanced by 900km 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 has some tendency 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 crustal scale models where the channel tunnels outward and is not exhumed by erosion (Beaumont et al. 2004, Fig.11a). The main difference from the crustal scale models is that the lower crust, with *B*(DMD/10)* does not subduct efficiently but instead tends to accumulate near the subduction point (Fig.10d). Unlike model LHO-1, where the lower crust forms a large antiform, the somewhat weaker and denser, eclogitic lower crust in LHO-LSI pools at the base of the isostatically depressed crust.

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 12My (Fig.11b). The slab is sufficiently dense that it subducts without significant deformation of the retro-mantle lithosphere and there is only a minor component of subduction zone advance. Between 18 and 21My 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 subduction zone retreats 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, sublithospheric mantle (Fig. 11c). This region widens to approximately 200km by 27My (Fig.11d) and temperatures in the upwelling fluid mantle range from 1000-1300°C. 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 overall mantle behaviour 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.11d) but in this case it is confined to the retro-side of the system. 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, however, similar to that of LHO-LS1. The difference is that 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 at one side of the orogen.

The two mantle-scale models illustrate how sensitive the behaviour of the mantle lithosphere may be to 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 such that 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 even when the behaviour of the lithospheric mantle is complex, and that the flows are not an artifact of the assumed basal boundary conditions in the crustal-scale models.

Discussion

Flow Modes in Temperature-Magnitude Space

Can we predict in which types of orogen the flow modes described here will operate? This can be answered in a general way using T-M diagrams (Fig.1) adapted to show where the flow modes are predicted in T-M space (Fig.12). The mode boundaries shown in Figure 12 are diagrammatic but can be refined if the T-M diagram is made more quantitative. In general, these flows do not occur in small cold orogens because normal quartzo-feldspathic crust is too strong. However, orogens that are rich in calcite and evaporite (e.g. anhydrite and halite) lithologies, which are much weaker than quartz-dominated lithologies, will develop these flows in the small-cold parts of T-M space. The positions of the mode boundaries in T-M space therefore vary with composition. It is, therefore, not surprising that the flow modes we have described are common in passive margin salt tectonics, even though passive margins do not have a large magnitude and are very cold.

Homogeneous channel flows are restricted to the hot regions of T-M space and their lower limit (Fig. 12) reflects a tradeoff between increasing temperature, which produces an increasingly thicker, viscously weaker, channel, and increasing magnitude, which amplifies the gravitational-driving force. However, no matter the magnitude, the crust must be hot and very weak. Volcanic arcs, as noted earlier, may be sufficiently hot and weak that channel flows can occur in restricted regions of the crust even at relatively small magnitude.

The tectonically-driven Hot Fold Nappes mode can occur in a much larger part of T-M space, including the Homogeneous channel domain (Fig. 12). All that is required is a sufficiently hot and thick orogen interior that nappes will be 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 purely mechanical Vise model, which was applied to the

Newfoundland Appalachians (Ellis *et al*.1998). Vise-type deformation occurs in the region of T-M space that is subcritical for the formation of Hot Fold Nappes (Fig.12). In this part of T-M space, the crustal Vise jaws are sufficiently strong and occupy most of the crust, not just the lower crust. Therefore, the orogenic crust is forced to contract and thicken when squeezed between the jaws of the Vise.

The Heterogeneous channel flow mode requires both tectonic forcing, to activate and evacuate weak lower crust, and gravitational forcing for the channel flow. The fully developed flow mode is therefore restricted to the Homogeneous channel flow region of T-M space because it requires the gravitational component (Fig.12). However, the tectonic evacuation of weak lower crust alone can occur in a larger part of T-M space (Fig. 12).

The Effect of Thermal Relaxation and Incubation Time on Crustal Flows

It helps to define and compare three timescales in order to explain the development of the flow modes seen in the model experiments. We define the incubation time to be the lag time between tectonic thickening of the crust and subsequent external processes, such as indentation or erosion, which act on the system. This definition is a generalization of the one used by England & Thompson (1984) that focused on erosion as the external process. The incubation time becomes important when compared with the time taken for internal processes, specifically in this case for radioactive heating and thermal relaxation, to achieve a particular thermal-rheological state within the crust. Whether external processes occur on timescales that are short or long with respect to internal process timescales determines the system response, as we describe in the examples below.

The behaviour of the Hot Fold Nappes model indicates one choice of an internal timescale. When viewed from the perspective of the stationary central part of the orogen, the strong lower crust indentor in model LHO-3 progressively underthrusts the orogen, forces the creation of the fold nappes, and expels them into the mid-crust (Fig. 7). This behaviour is typical of models in which the crust is sufficiently weak to deform under gravitational forces, as demonstrated by the development of a topographic plateau in the interior of the model (Fig. 7, a and b, 30-40My). Plateau development demonstrates that the crust cannot sustain thickness variations against gravitational forces and has flowed to equilibrate the pressure in the crust below the plateau. Under these circumstances, the crust approaches a hydrostatic state and flows over the lower-crustal indentor because the indentor cannot force thickening of the weak crust against its weight. This near-hydrostatic response contrasts with that of strong crust, which shortens and thickens when indented because its response is not limited by its weight. The mechanical response to indentation therefore depends on the thermal-rheological state of the model crust.

We therefore define the threshold time, τ_{HN}, as the delay time necessary for thermal relaxation to achieve the thermal-rheological state required for the hot nappe type of response seen in LHO-3. If the incubation time > τ_{HN}, indentation will create hot nappes and the converse. 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. For example, thickened cold crust with low levels of radioactive heating will thermally relax with a certain timescale but it may never become weak enough for the hot nappe response, giving the system a finite thermal relaxation timescale but an infinite $\tau_{HN.}$ Conversely, crust that is already hot and weak before thickening can have a τ_{HN} that is only a fraction of the thermal relaxation timescale. Model LHO-3, with lateral injection of hot fold nappes, corresponds to a system for which the incubation time was equal to or greater than τ_{HN}. For natural orogens we lack details of the thermal-rheological evolution; without this information the development of an orogenic plateau provides a minimum measure that the threshold timescale τ_{HN} has been exceeded.

Model LHO-1 illustrates the analogous situation in regard to the onset of channel flows. We define τ_{CF} as the threshold delay time for the onset of channel flow. If the incubation time > τ_{CF} channel flow will develop. Given that channel flow requires weaker crust than that for plateau development, $\tau_{CF} > \tau_{HN}$ for the same type of crust. This relationship explains the LHO-3 behaviour in which the initial response to indentation is to produce fold nappes (Figs. 7 and 8, a and b, 50My), that is incubation time > τ_{HN} and < τ_{CF}, but a later channel flow is superimposed on the system (Fig. 7a, 65My) demonstrating that the incubation time is now > τ_{CF}.

Model LHO-2 also illustrates the effects of incubation. During deformation phase one, the *B*(DMD/10)* regions in the lower crust are detached and evacuated as they are incorporated into the model orogen (Fig. 5b, 30 and 50My). These regions are already sufficiently hot and weak that they require no incubation in order to flow horizontally. In contrast, the overlying mid-crust deforms by shortening and thickening, the characteristic response of strong crust. The gravitationally driven large-scale channel flow (Fig. 6, a and b, 50-60My) develops only during the diachronous second phase, following incubation of the system, demonstrating that the incubation time is now > τ_{CF}.

In summary, the flow modes can be predicted by comparing the incubation time with the two threshold delay times defined above. However, the behaviour can be complex because the system evolves diachronously, such that the incubation time increases from the exterior to interior regions. By implication different flow modes can coexist in different regions of the model.

The Flow Modes Ternary Diagram

Given the potential complexity of the flow modes it useful to consider end member circumstances for which Homogeneous, Heterogeneous and Hot Fold Nappe styles will and will not occur. Flow mode space can be represented by a ternary diagram (Fig.13) with the three modes considered here as end members. The flow mode space is divided into two regions where gravitationally and tectonically driven flows will operate. The Hot Fold Nappe end member requires that τ_{HN} < incubation time < τ_{CF}, that is, the system is too strong for gravitationally-driven channel flows. Homogeneous channel flows occur when $\tau_{HN} < \tau_{CF}$ < incubation time but there is no external tectonic forcing, for example by indentors. Heterogeneous channel flows are initially tectonically driven but can evolve diachronously to include a superimposed channel flow. Although the flow modes ternary diagram does not consider all possibilities, pathways within the diagram are a useful way to consider the evolution of modes in natural or hypothetical orogens. For example, the LHO-3 pathway leads from the Hot Fold Nappes end member into the gravitationally-driven region, and thence to the Heterogeneous channel flow end member.

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 (Culshaw *et al.* submitted, and references therein). The main difference among the models is in the cause and timing of the development of the Infrastructure.

In model LHO-3 (Fig. 7) the orogen develops an Infrastructure of lower and mid-crustal nappes partly overlying the underthrust indentor and decoupled from the upper crust by a reverse sense shear at the top of the highly strained ductile mid-crust (Fig. 7,a and b, 50-65My). The upper crust, the Superstructure, remains relatively undeformed after its Phase 1 shortening except where there is syntectonic exhumation by erosion (compare Fig. 7a and 7b). In contrast, Phase 1 structures in the mid- and lower-crustal Infrastructure are strongly overprinted by Phase 3 flow (Fig. 7a and b, 50-65My). From an observational perspective, the three-phase evolution of model LHO-3 leads to the development of 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 Phase 3 younger deformation activated by the collision with the indentor. In this model the mid- and lower-crust become weak on a τ_{HN} timescale of approximately 20My, but this region is not strongly activated and deformed until the indentor collides much later. The Infrastructure is therefore created by the tectonic process, not an internal gravitational flow, after a much longer incubation time. In contrast models LHO-1 and LHO-2 develop a sufficiently weak Infrastructure during incubation that it deforms and flows under gravitational forces alone (Figs. 3 and 5). Under these circumstances Superstructure/Infrastructure relationships develop on the threshold delay timecale τ_{CF} for channel flows and do not require external forcing by, for example, an indentor or other tectonic processes.

Conclusions

We draw four main conclusions.

1) Gravitationally driven mid-crustal channel flows, exemplified by the Homogeneous or Heterogeneous modes (Fig. 14), are most likely end members that will occur in Giant and Supergiant members of the Large Hot Orogen family. Such flows are considered likely to occur in the Himalayan-Tibetan system. Gravitationally driven flows may have been more prevalent in Archean orogens, if they were indeed hotter than equivalent sized contemporary orogens.
2) Other tectonically forced flow modes, exemplified by the Hot Fold Nappes mode and the tectonic component of the Heterogeneous Flow mode (Fig. 14), may occur in Giant and Supergiant orogens (Fig. 12). More importantly, they can occur in Large Hot Orogens that are subcritical with respect gravitationally driven flows. In particular, the Hot Fold Nappes mode is predicted for accretionary and collisional orogens where the orogen involves collision/indentation by strong crust, for example, older second or higher cycle refractory crust such as a cratonic nucleus, or cold oceanic crust.
3) 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 position in T-M space and that this approach offers a way to predict the types of possible crustal flow modes.
4) Elsewhere, Jamieson *et al*. (2004) interpret the Grenvillian orogen in Canada to record the diachronous evolution of Hot Fold Nappes with superimposed Heterogeneous Channel Flows during its collision with Laurentia. We suspect that these tectonically driven styles will be recognized in many North American orogens owing to the particular style of successive collisions against and accretion to, the cratonic core. In particular, the Trans Hudson, central Appalachian,

southern Canadian Cordillera orogens, and the Archean Slave and Superior cratons are prime candidates for these flows.

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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.

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 (**a**) pro crust and (**b**) 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.4. Model LHO-1 results showing evolution of the velocity field (horizontal lines are velocity vectors) and temperature (isotherms shown at 100°C intervals) 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.5. Model LHO-2 results showing evolution of crustal-scale deformation for (**a**) pro crust and (**b**) 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.6. Model LHO-2 results showing evolution of the velocity field (horizontal lines are velocity vectors) and temperature (isotherms shown at 100°C intervals) 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.7. Model LHO-3 results showing evolution of crustal-scale deformation for (**a**) pro crust and (**b**) 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-3 results showing evolution of the velocity field (horizontal lines are velocity vectors) and temperature (isotherms shown at 100°C intervals) 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.9. Configuration and principal properties of the upper mantle scale models, LHO-LS1 and LHO-LS2. These models are the same except for the minor difference in the reference densities of their respective mantle lithospheres. Notation $\phi = 15° \rightarrow 2°$ 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 by 10); *B*(WOl×10)* (Wet Olivine rheology scaled by 10); ρ = density given at reference temperatures; thermal coefficient of volume expansion =3x10^{-5}. 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 120 km and sublithospheric mantle. Lithosphere converges asymmetrically from right 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 right is the initial narrow weak zone. There are no surface processes. Bold frame shows area displayed in Figures 10 and 11; note that the position of this frame migrates with time in these figures.

Fig.10. Model LHO-LS1 results showing evolution of upper mantle-scale deformation. Panels show the model materials (see Fig. 9), a sparse version of the Lagrangian tracking grid, the velocity field (arrows, scale at bottom), and isotherms contoured at 100°C and 50°C intervals in the lithosphere and sublithospheric mantle, respectively (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 30My.

Fig.11. Model LHO-LS2 results showing evolution of upper mantle-scale deformation. Panels show the model materials (see Fig. 9), a sparse version of the Lagrangian tracking grid, the velocity field (arrows, scale at bottom), and isotherms contoured at 100°C and

50°C intervals in the lithosphere and sublithospheric mantle, respectively (V = H). t = elapsed model time, Δx = total convergence. No surface processes. In this case the 200 km movement of the last panel window is to the right. Crustal channel flow is restricted to the retro-crust but is well developed by 30My.

Fig.12. Interpretation of flow modes discussed in this work in regard to the positions where they may operate in T-M space.

Fig.13. Flow modes ternary diagram designed to emphasize the continuum nature of the modes and the relative importance of gravitational and tectonic forcing within this continuum. $\eta_C = 10^{19}$ Pa.s is regarded as the critical viscosity for the mid-crust to develop gravitationally-driven channel flows, so that, $\eta_{eff} > \eta_C$ defines the tectonically-driven flow modes regime and, $\eta_{eff} < \eta_C$ defines the gravitationally-driven flow modes regime.

Fig.14. 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.