# Crustal Flow Modes in Large Hot Orogens: Appendix

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**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.

## 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 $\phi_{eff} = 15°$ 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 ($K$, $\kappa$, $C_P$, $\rho$) 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 $\phi_{eff} = 5°$, representing sedimentary rocks of the upper crust with high pore fluid pressures. The middle crustal layer (10-25 km) uses *B*(WQx5)* with $\phi_{eff} = 15°$, 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 $\phi_{eff} =$ 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 $h^3/\eta$; 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.