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Thrust Belts and Foreland Basins

From Fold Kinematics to Hydrocarbon Systems

With 290 color figures

Springer

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Preface

Thrust belts and foreland basins record the main phases of orogenic evolution. They are shaped by the coupled influence of deep (flexure, plate rheology and kinematics) and surficial (erosion, sedimentation) geological processes, at different time scales. Thrust belts and foreland basins constitute important targets for scientists interested in both fundamental and applied (fluids, hydrocarbons) aspects.

In the framework of a new cycle of workshops of the ILP task force on "Sedimentary Basins", a three-day meeting on "Thrust Belts and Foreland Basins" was organized in December 2005 on behalf of the Société Géologique de France and the Sociedad Geológica de España, hosted by the Institut Français du Pétrole near Paris. The main purpose of the meeting was to offer the opportunity for Earth scientists from different disciplines, i.e. geologists, geophysicists and geochemists, to present and share their different knowledge on the processes governing the evolution of orogenic belts and adjacent forelands. A special emphasis had been given to make a "bridge" between the most recent advances in surface processes, geochemistry, provenance studies, field studies, analogue and numerical modelling, high resolution seismicity, and hydrocarbon prospect in forelands basins. The conference was successful in bringing together scientists from academia and industry from nearly 20 countries. New contributions using the geologic information recorded in thrust belts and foreland basins as well as stimulating key notes provided fertile ground for discussion, focusing on the orogenic evolution of adjacent mountain belts, on the stratigraphic records resulting from the coupled influence of deep and surficial geological processes, on exploration strategies for hydrocarbons in foothills areas, and on recent methodological and technical advances that have renewed our view on these important targets in both their fundamental and applied aspects.

The present volume addresses most of these topics. It comprises 25 key papers presented at the conference. The content of the volume reflects the diversity of the presentations and the success of the workshop, and is likely to promote new contacts between interdisciplinary Earth scientists. Volume architecture brings the reader from geodynamic considerations to general and specific issues of thrust belt description and hydrocarbon systems exploration through seismic imaging, fluid flow studies, and structural modelling. Given the focused attention that Zagros/Makran and Carpathian thrust belts received during the meeting and volume elaboration, contributions specific to these two important areas were put separately. The varied methodologies implemented when studying these two thrust belts examples, and the contrasted answers they bring, stress the importance of confronting independant approaches. These case studies also provide to any researcher interested by thrust belts a synthetic view on the modern techniques and recent advances developed for studying these major geological targets.

Olivier Lacombe, Jérôme Lavé, François Roure
and Jaume Vergés

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Olivier is an expert in brittle tectonics, especially in the study of fracture populations and reconstruction of paleostress orientations and magnitudes using analysis of faults and calcite twins. His main interest relates to the understanding of the structure and the kinematic evolution of fold-thrust belt - foreland basin systems, with special emphasis on the style of deformation (basement-involved tectonics and inversion), timing of deformation and tectonic history. Olivier also recently focused on the characterization of the ductile to brittle transition during late stage exhumation of HP metamorphic rocks in orogen hinterlands. He has worked in various regions worldwide, i.e., in the Alps and Pyrenees-Provence, Taiwan, Zagros, and more recently in the Cyclades-Aegean.

Dr. Jérôme Lavé is research scientist at the CNRS (Centre National de la Recherche Scientifique) in the LGCA (Laboratoire de Géologie des Chaînes Alpines) in Grenoble (France) since 1999. He obtained his PhD in geophysics from the Denis Diderot University (Paris, France) in 1997.

Jérôme is an expert in seismotectonics and morphotectonics. He has been working on terrace deformation in relation to active folding in thrust belts, on erosion and river incision processes and on quantification of pebble and bedrock abrasion processes during fluvial transport. He recently developed a new paleo-altimetric method based on the cosmogenic nuclides. His general fields of interest are the coupling and feedback loops between tectonic, erosion and climate, the dating and modelling of fluvial and glacial landscapes in relation with paleoclimate, and the numerical modelling of landscape evolution. He has mostly worked in the Himalayas, Tibet, Zagros and Altiplano.

Dr. François Roure is currently professor at IFP (Institut Français du Pétrole) and VU-Amsterdam. He joined the CNRS (Centre National de la Recherche Scientifique) in 1980. After a PhD aiming at the study of the Mesozoic geodynamic evolution of the North American Cordillera in N California and Oregon, he moved to IFP in 1984, studying first the crustal architecture of the Alps, Pyrenees and Apennines, then the petroleum systems in foothills and foreland domains in the Carpathians, Albanides, Sicily, Alaska and Colombia.

More recently, François has focused his interest on the evolution of sub-thrust reservoirs in the Himalayan foothills in Pakistan, as well as in the sub-Andean basins, and the North American Cordillera again, i.e., in Mexico and in the Canadian Rockies. His last projects aimed at the study of the crustal architecture and petroleum evaluation of the United Arab Emirates and North Algeria.

Dr. Jaume Vergés is researcher at the Institute of Earth Sciences "Jaume Almera" of the CSIC (Consejo Superior de Investigaciones Científicas) in Barcelona (Spain). He has been working on many fold-and-thrust belts, i.e., the Pyrenees, Zagros, Andes, NW Cuba and the Polish Carpathians, mostly on geometries of folding and thrusting and timing of deformation.

Jaume is presently co-director of the GDL (Group of Dynamics of the Lithosphere), a working group focusing on understanding the links between surficial and deep geodynamic processes through combination of geology, geophysics and numerical modelling.

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Part I
Surficial and Deep Processes in Thrust Belts

Part I of the volume comprises two chapters which deal with several large-scale and first-order features of thrust belts and foreland basins.

Burov (Chapter 1) investigates the interactions between the surface and subsurface processes by means of thermo-mechanical modelling. One main point is that advection of material at the Earth's surface and horizontal flow in the lower crust might be coupled so as to permit mountain growth in response to horizontal shortening. This mechanism is investigated on the basis of semi-analytical and numerical experiments in which the rheological layering of the lithosphere and surface processes are modelled. These ideas are tested on well-studied cases such us the Western Alps, the Tien Shan and the Himalaya. Some implications about the role of climate on continental tectonics and on the geomorphology of mountain ranges are derived.

In their paper, Lenci and Doglioni (Chapter 2) address the overall asymmetry of thrust belts by analyzing various convergent margins in terms of geographic polarity of the margin, age and composition of the subducting plate. They argue that the asymmetry between orogens or accretionary wedges is to the first-order global and related to geographically opposed (i.e., W- to E-NE directed) subduction zones, while local/regional stratigraphic-rheological characteristics which may vary along strike, such as the décollement depth, exert only second-order controls on each orogen

Coupled Lithosphere-Surface Processes in Collision Context

Evgueni Burov

Abstract. From the mechanical point of view, a mountain range that exceeds a certain critical height (of about 3 km in altitude, depending on rheology and width) should flatten and collapse within few My as a result of gravitational spreading of its ductile crustal root. Even if the crustal root does not collapse, the mountain range would be levelled by gravity sliding and other surface processes that, in case of static topography, lead to its exponential decay with a characteristic time constant on the order of 2.5 My. However, in nature, mountains grow and stay as localized tectonic features over geologically important periods of time (> 10 My). To explain the paradox of long-term persistence and localized growth of the mountain belts, a number of workers have emphasized the importance of dynamic feedbacks between surface processes and tectonic evolution. Indeed, surface processes modify the topography and redistribute tectonically significant volumes of sedimentary material, which acts as vertical loading over large horizontal distances. This results in dynamic loading and unloading of the underlying crust and mantle lithosphere, whereas topographic contrasts are required to set up erosion and sedimentation processes. Tectonics therefore could be a forcing factor of surface processes and vice versa. One can suggest that the feedbacks between tectonic and surface processes are realized via two interdependent mechanisms:

1. Slope, curvature and height dependence of the erosion/deposition rates
2. Surface load-dependent subsurface processes such as isostatic rebound and lateral ductile flow in the lower or intermediate crustal channel.

Loading/unloading of the surface due to surface processes results in lateral pressure gradients, that, together with low viscosity of the ductile crust, may permit rapid relocation of the matter both in horizontal and vertical direction (upward/downward flow in the ductile crust). In this paper, we overview a number of coupled models of surface and tectonic processes, with a particular focus on 3 representative cases:

1. Slow convergence and erosion rates (Western Alpes)
2. Intermediate rates (Tien Shan, Central Asia)
3. Fast convergence and erosion rates rates (Himalaya, Central Asia).

1 Introduction

Continental mountain belts, such as, for example, Tien Shan (Central Asia, Figure 1a), are characterized by highly localized topography elevations persistently growing over tens of millions of years. The fact that gravitational potential energy per unit surface $0.5\rho g h^2$ scales as h^2 implies that a thrust belt should grow more easily in width than in height (Molnar and Lyon-Caen, 1988, h is the mean topography elevation above sea level, ρ is density and g is acceleration due to gravity). A portion of continental crust submitted to quasi-static horizontal shortening should tend to thicken homogeneously. This can be put another way around by considering that a range results from thrusting on faults that cut through the upper crust and root into the lower crust. Uplift of the range implies an increase in the vertical stress acting on the fault. This acts to oppose further frictional sliding on the fault, inhibiting further thrusting. A new fault will then form farther away from the range front leading to widening of the range. In addition, erosion and sedimentation at the surface, together with flow in the lower crust, should favor smoothing of topographic irregularities. At the pressure and temperature conditions of the lower crust, most crustal rocks are thought to flow easily at very low deviatoric stresses (e.g., Brace and Kohlstedt, 1980; Wang et al., 1994, Fig. 2a). The deviatoric stresses associated with slopes of the topography and of the Moho (e.g., Fleitout and Froidevaux, 1982) should therefore be relaxed by viscoplastic flow in the ductile lower crust inducing decay of topographic irregularities (Kusznir and Matthews, 1988; Gratton, 1989; Bird, 1991, Fig. 1b).

The growth and maintenance of topographic features at the surface of continents might be taken to indicate that the strength of the crust exceeds the deviatoric stresses associated with slopes of the topography and of the Moho. Yet, as mentioned above, laboratory experiments indicate that at the pressure and temperature conditions of the lower crust, most crustal rocks should flow easily. Irregularities of the topography and of the Moho boundary should therefore be

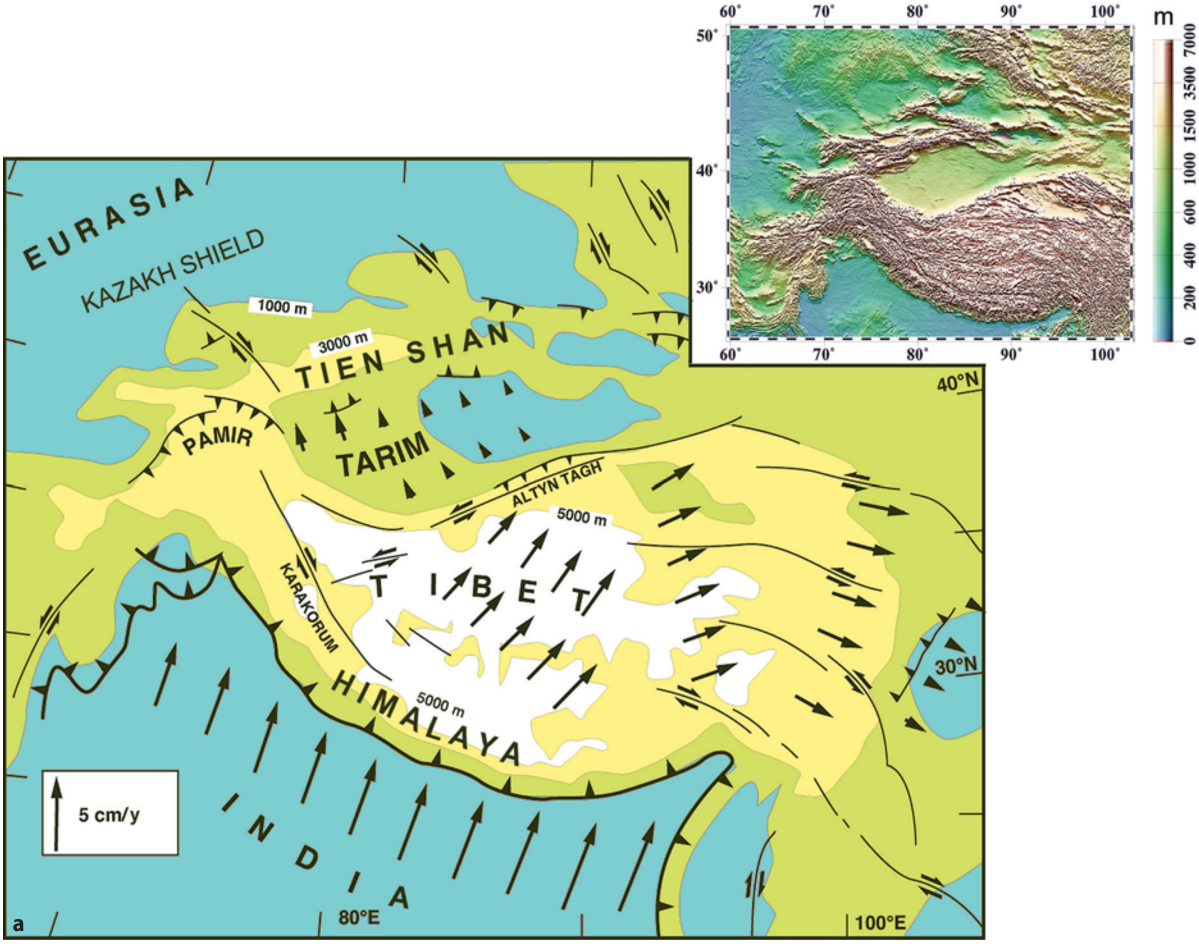


Fig. 1a. Actively growing intercontinental belts and plateaux: an example showing a schematic map of India-Eurasia collision with its main features such as the Himalayan mountain belt, Tibetan plateau, Tarim basin, Pamir and Tien Shan mountain belt. Insert shows a digital elevation map of the same area. The topography peaks to 8800 m in the Himalayas (Everest) and 7500 m in Tien Shan (Pobeda Peak). Modified after (Avouac and Tapponier, 1993)

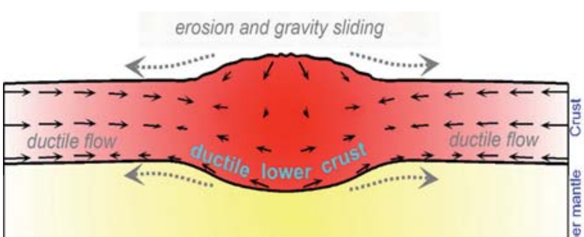
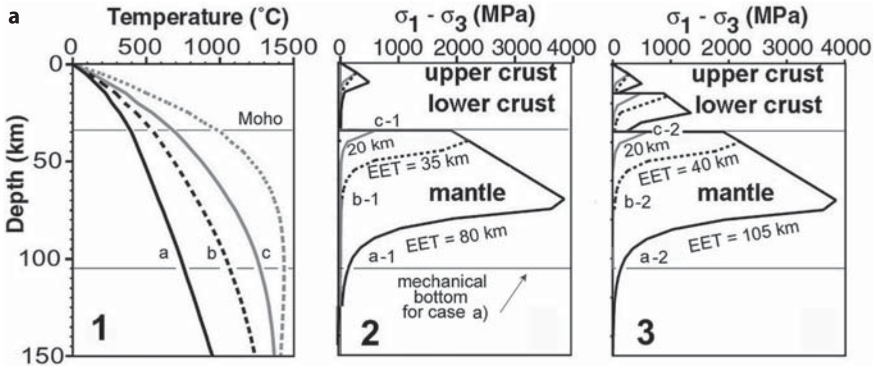


Fig. 1b. Erosional and gravity collapse of a mountain range (e.g., Gratton, 1989; Bird, 1991). In the conceptual model shown here, there is no balance between surface and subsurface processes. Even if such range was created somehow, it will not persist, as its root and topography will be flattened in about 2 My in the absence of some compensating mechanisms. Tectonic convergence may not solely compensate this flattening; it may only grant an overall thickening of the crust; some additional localizing mechanisms are needed to concentrate thickening in a narrow range.

relaxed by viscoplastic flow in the ductile lower crust and decay with time (Kusznir and Matthews, 1988; Gratton, 1989; Bird, 1991). Consider, for example, the Tien Shan range, which is, except for the Himalayas, the largest and most active intracontinental range in the world (Fig. 1a). Tien Shan (translated as "Heavenly Mountains") is 300–400 km wide in its central area, with a mean elevation of about 3500 m and local peaks of up to 7500 m, in a zone of relatively thick and tectonized crust (Moho depths from 50 to 70 km) (e.g., Avouac et al., 1993). The Tien Shan is a continuously growing range, that has started to rise 10–15 My ago. A simple dimensional analysis (Gratton, 1989) as well as numerical simulations (Bird, 1991; Avouac and Burov, 1996; Burov and Watts, 2006) show that the topography of such a range should, instead of growing, be reduced by half in a few My (Fig. 1b). This estimate is based on the assumption of ductile rheology of the



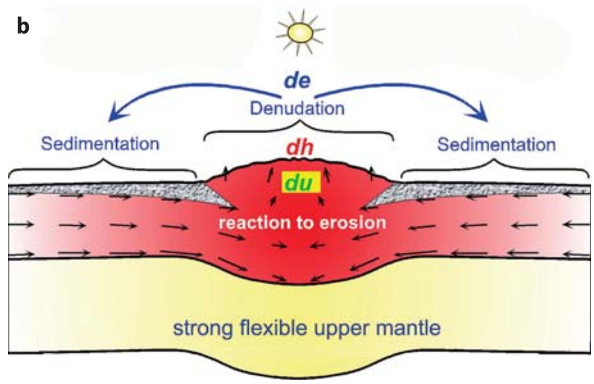


Fig. 2. **a**) Typical rheology profiles for continental lithosphere indicate the possibility for lower-crustal flow: (*1*) geotherms that yield YSEs shown in the middle and on the right; (*2*) yield stress envelope (YSE) for quartz-rich upper and lower crust and olivine mantle; (*3*) yield stress envelope (YSE) for quartz-rich upper crust , diabase lower crust, olivine mantle. EET – equivalent elastic thickness of the lithosphere computed for each of YSEs. **b**) Solution to the problem stated in Fig. 1b: a conceptual model of continental collision in which strong feedback between surface processes, isostatic reaction and subsurface crustal flow results in accelerated growth of topography in the area of strongest subsurface uplift

lower crust, which is supported for this area by multiple data starting from seismic data (Vinnik and Saipbekova, 1984; Makeyeva, 1992; Roecker et al., 1993; Vinnik et al., 2006) and ending by gravity-flexural analysis (Burov et al., 1990, 1993; Avouac and Burov, 1996). Only the short topographic wavelengths, typically less than a few tens of kilometers that can be supported by the strength of the upper crust would be maintained over geological periods of time, yet provided that they are not wiped out by erosion, which is faster on short wavelength features. In addition, surface processes might be thought to contribute to an even more rapid smoothing of the topography. Similarly, in the absence of strong rheological heterogeneities or of strain localization processes, a portion of a continental crust submitted to horizontal shortening should tend to thicken homogeneously, so that no mountain should form. The growth and maintenance of an intracontinental mountain range over long periods of time must therefore involve dynamical processes allowing for long-term localization of lithospheric strain below the mountain.

Several mechanisms have been advocated to explain localization of major thrust faults and, by its proxy, stability of mountain belts. Intrinsic strain softening properties of rocks could sustain localized thrust faulting at the crustal scale. Alternatively, a range could result from shear stresses at base of the crust induced by lithospheric under-thrusting or by mantle dynamics (e.g., Beaumont et al., 1994; Ellis et al., 1995). Such a mechanism may be suggested for mountains associated with subduction zones or with hotspots (Vogt, 1991), but seems inappropriate to explain most intracontinental mountains. In the case of, for example, the Tien Shan belt, a particular mantle dynamics has been inferred from gravity modelling (Burov et al., 1990, 1993) and seismic anisotropy (Makeyeva, 1992; Roecker et al., 1993), but we contend that it might not be the key factor. Our point is instead that coupling between surface processes and flow in the lower crust could provide an alternative and more general explanation (Avouac and Burov, 1996; Burov and Cloetingh, 1997).

To explain the paradox of long-term mountain persistence and localized growth, a number of workers have emphasized the importance of dynamic feedbacks between surface processes and tectonic evolution (e.g., Molnar and England, 1990; Masek et al., 1994a; Avouac and Burov, 1996; Molnar, 2001). Indeed, surface processes modify the topography and redistribute tectonically significant volumes of sedimentary material (vertical, or normal loads) over large horizontal distances. This may result in dynamic loading and unloading of the underlying crust and mantle lithosphere, whereas topographic contrasts are required to set up erosion and sedimentation processes.

Tectonics therefore could be a forcing factor of surface processes.

In this paper, we first review the existing models of surface processes and the thermo-mechanical properties of the lithosphere that condition its response to surface and tectonic loading-unloading. We then review our own and other previous modelling studies that show that surface and tectonic processes are not independent processes and can interact. We show in particular that advection of material at the Earth's surface and horizontal flow in the crust might be coupled so as to permit mountain growth in response to horizontal shortening. This mechanism is then validated and investigated on the basis of semi-analytical and numerical experiments in which the rheological layering of the lithosphere and surface processes are modelled. We then find that, depending of the erosion rate compared to horizontal shortening, flow in the lower crust can be "outward" (from under the high topography) or "inward" (toward the crustal root of a high topography). When inward flow occurs, a mountain range can actually grow and no other mechanism is required to explain localized uplift. Some implications about the role of climate on continental tectonics and on the geomorphology of mountain ranges are then derived.

We suggest an additional feedback mechanism by lateral crustal flow (Fig. 2b). According to this mechanism, erosional removal of material from topographic heights (dynamic unloading) and its deposition in the foreland basins (dynamic loading) should result in horizontal ductile crustal flow that may oppose gravitational spreading of the crustal roots and may eventually drive a net influx of material towards the orogeny. We finally test our ideas on three representative and well-studied cases:

1. Slow convergence and erosion rates (Western Alpes),
2. Intermediate rates (Tien Shan, Central Asia), and
3. Fast convergence and erosion rates rates (Himalaya, Central Asia).

2 Interplays Between Surface and Tectonic Processes

2.1 Tectonic Forcing on Surface Processes

Surface topography elevations are required to set up erosion and sedimentation processes. Tectonics is therefore a forcing factor of surface processes. Following Ahnert (1970), and Pinet and Souriau (1988), Summerfield and Hulton (1994) have compiled rates of denudation at the scale of major river basins. These studies indicate that denudation is primarily influenced by basin topography so that rates of denudation appear to be systematically high in areas of active tectonic uplift. Common values of mean denudation rates in such areas would be of the order of a few 0.1 mm/y to about 1mm/y at the scale of large drainage basins. Such rates are generally consistent with estimates derived from balancing sediment volumes over geological periods of time (Leeder, 1991; Summerfield and Hulton, 1994). Thermochronologic studies indicate, however, local values as great as 1 mm/y (see Leeder, 1991 and Molnar and England, 1990, for critical reviews). The discrepancy between local and basin averaged estimates is due to the fact that tectonic uplift is probably distributed in brief pulses over localized domains within a drainage basin (Copeland and Harrison, 1990). In the absence of any tectonic feedback, common values of denudation rates should lead to the disappearance of a major mountain belt like the Alpes or Tien Shan in a few million years. Pinet and Souriau (1988) demonstrated that denudation leads to an exponential decay of the topography of a range with a characteristic time constant of the order of 2.5 m.y.

2.2 Coupling Between Denudation and Tectonic Uplift due to Isostasy

Many recent models have investigated coupling between the isostatic reaction and surface processes (e.g., Kooi and Beaumont, 1994; Snyder et al., 2000; Basile and Allemand, 2002; Garcia-Castellanos, 2002; Garcia-Castellanos et al., 2002; 2003; Simpson and Schlunegger, 2003; Persson et al., 2004; Casteltort and Simpson, 2006). Redistribution of surface loads by erosion and sedimentation must induce tectonic deformation to maintain isostatic balance. Vertical uplift is expected to partly compensate unloading in the area subjected to denudation while subsidence should occur in response to loading by sedimentation. This feedback mechanism may lead to some coupling between denudation and tectonic uplift (e.g., Ahnert, 1970). A first consequence is that the time needed to erode a topographic relief must take into account removal of the topographic relief and of the crustal root. If local isostasy is assumed and if horizontal strains are neglected, denudation is dynamically compensated by uplift and the characteristic time of decay of the topography would then be of the order of 10 m.y. (Leeder, 1991). In addition, it has been argued that a positive feedback may arise (Molnar and England, 1990; Masek et al., 1994b). If the slopes of valleys steepen during river incision, isostatic readjustment following denudation in a mountain range may result in a net uplift of the higher summits in spite of the average lowering of reliefs. Alternatively regional compensation due to the elasticity of the lithosphere might lead to the uplift of the eroded edge of a plateau. Erosion might therefore induce some uplift of topographic summits leading in turn to enhanced erosion. The uplift of the Himalayan belt during the last few million years may have resulted from such a coupling rather than from thrusting at the Himalayan front (Burbank, 1992; Burbank and Verges, 1994). Note however that, while the peaks might reach higher elevations following isostatic adjustment, the net effect of erosion is crustal thinning. Thus, these models cannot explain the growth of mountains over long time periods.

The strongest feedback between erosion and isostatic reaction would be obtained for local isostasy. It will be mitigated in case of more regional compensation and become negligible for lithospheres whose equivalent elastic thickness exceeds 60 km. This is another reason to support the idea that more efficient mechanisms should also take place in collisional settings.

2.3 Coupling Between Surface Processes and Horizontal Strains

As mentioned in the introduction, small lateral variations of the crustal thickness should drive horizontal flow in the lower crust. Some studies have already pointed out to the importance of such a process in continental tectonics (e.g., Lobkovsky, 1988; Lobkovsky and Kerchman, 1991; Burov and Cloetingh, 1997). For example, Kruse et al. (1991) have shown that horizontal flow in the lower crust has regulated isostatic equilibrium during extension in the Basin and Range. The lower crust would have been extruded from under the high topography during that process. Following Westaway (1994) we will call this sense of flow "outward". On the other hand, Gregory and Chase (1994) inferred "inward" flow, toward the crustal root, during the Laramide orogeny of the Frontal Range, Colorado. The characteristic time associated with flow in the lower crust induced by the topography of a range a few thousands of meters high, a few hundreds of km wide, is in the order of a few m.y. The characteristic times of erosional decay of the topography of a range and of lateral collapse of a crustal root are thus of the same order of magnitude. Since both processes are driven by topographic slopes, some coupling may arise. Although it is not often pointed out, it has long been recognized that this kind of process might play a major role in elevation changes within continents (see Westaway, 1994 for an review of historical development of these ideas). Westaway (1994) made a case for such a coupling, with inward flow, in the context of extensional tectonics in western Turkey. He proposed that sediment loading in the sedimentary basins would have driven flow toward the uplifted area. This kind of process was first modelled by King and Ellis (1990), who modelled crustal extension using a thin elastic plate (upper crust) overlying an inviscid fluid (lower crust).

We propose that this kind of coupling might also appear in a compressional context. Let us consider a portion of a lithosphere, loaded with some initial range topography in regional isostatic balance, and submitted to horizontal compression. Horizontal stress gradients, resulting from the slopes of the topography and of the Moho, must drive horizontal flow. The lithosphere in the region of the range is weakened, since the crust is thick and hot, and because bending of the lithosphere beneath the mountain load tends to reduce its strength (Burov and Diament, 1992; 1995; Ranalli, 1995). Higher strain rates in the area below the range should therefore be expected. A low viscosity channel in the lower crust beneath the high topography might therefore allow lateral flow. In the absence of horizontal shortening and erosion, the lower crust below the range would be extruded laterally as discussed by Bird (1991). If erosion takes place, a regime may be established in which horizontal shortening would be preferentially accommodated by crustal thickening in the area below the range:

a) Surface processes remove material from the range and feed the adjacent flexural basins inducing isostatic imbalance.
b) This imbalance produces a temporary excess of normal stress below the foreland basins and deficit below the range favoring flow in the lower crust towards the crustal root. The range uplifts and the basins subside.

Ultimately this coupled regime might lead to some dynamic equilibrium in which the amount of material removed by erosion would balance the material supplied to the range by subsurface deformation.

Apart of the direct mechanical effect of erosion/sedimentation (loading-unloading) on the lithosphere, it also has very important thermal, and, by proxy, mechanical consequences, because the removal and accumulation of sedimentary matter modifies surface heat flux and thermal conditions in the upper crust (e.g., England and Richardson, 1977). Accumulation of sediments in the forelands leads to (1) cooling of the accretion wedge at a short term, in case of rapid advection/filling (initial stages of collision when the convergence rate is highest); (2) heating of the accretion wedge at a long term in case of slow advection (when collision rate slows down), due to heat screening (sediments have low thermal conductivity) and the abundance of heat producing radiogenic elements in the sedimentary matter. Furthermore, penetration of the mechanically weak sediment in the subduction channel should serve as lubrification and may enhance the conditions for subduction processes.

2.4 Coupling of Surface Processes and Tectonic Input/Reaction in Full Scale Mechanical Models: Major Stages

A number of earlier modelling studies (e.g., Beaumont, 1981; Beaumont et al., 1992; 1995; Willet, 1999) have investigated various relationships between erosion and tectonic processes. However, tectonic reaction was not fully accounted for, as most of these models that have exploited semi-kinematic formulations for the crust or the mantle lithosphere. One of the first full-scale parametric semi-analytical models was developed by Avouac and Burov (1996) in order to validate the coupled regime between surface and subsurface processes. For this purpose this model accounted for:

1. Surface processes.
2. The effect of topographic loads and variations of crustal thickness on the mechanical behavior of the lithosphere.
3. Ductile flow in the lower crust.
4. Depth-and-strain dependent rheology of the lithosphere.

In the following sections we first discuss the components needed to build a coupled tectonic model of orogenic building:

1. The existing models of surface processes.
2. The rheology data needed for proper account of the mechanical response of the lithosphere.
3. Thermal models of the lithosphere needed for proper account of thermally dependent ductile rheology.

We then describe the design and major results of the coupled semi-analytical model of Avouac and Burov (1996). This semi-analytical model has a number of limitations in terms of model geometry and its inability to account for some key deformation modes such as formation of major thrust faults. For this reason, in the final sections of this study, we go further by introducing an unconstrained fully coupled numerical thermo-mechanical model of continental collision/subduction similar to that used by Burov et al., (2001); and Toussaint et al. (2004a,b). This model takes into account more realistic (than in the previous studies) geometry of the convergent plates, accounts for large strains and brittle-elastic-ductile rheology including localized brittle (faulting) and ductile deformation.

3 Surface Processes Modelling: Principles and Numerical Implementation

3.1 Basic Models of Surface Processes

A growing amount of field and experimental studies have investigated and validated various forms of long-and-short range erosion and sedimentary transport laws and models (Ahnert, 1970; Beaumont, 1981; Beaumont et al., 1992;2000; Burbank, 1992; Burbank and Verge, 1994; Ashmore, 1982; Mizutani, 1998; Lavé and Avouac, 2001; Lague et al., 2000, 2003; Davy and Grave, 2000; Lague et al, 2000; Molnar, 2001; Grave and Davy, 2001; Densmore et al., 1997;1998; Pinet and Souriau, 1988).

Short-range erosion. A simple two-dimensional law may be used to simulate erosion and sedimentation at the scale of a mountain range. The evolution of a landscape results from the combination of weathering processes that prepare solid rock for erosion, and transportation by hillslope and stream processes (see Carson and Kirkby, 1972 for a review). Although many factors, depending on the lithologies and on climate (e.g., Fournier, 1960; Nash, 1980), may control this evolution, quite simple mathematical models describing the geometrical evolution of the morphology at the small scale have been proposed and tested successfully (e.g., Kirkby, 1971; Smith and Bretherton, 1972; Chorley et al., 1984; 1986; Luke, 1972; 1974; Kirkby et al., 1993). For example, the two-dimensional evolution of a scarp-like landform can be modelled assuming that the rate of downslope transport of debris, q, is proportional to the local slope, ∇h (Culling, 1960; 1965; Hanks et al., 1984; Avouac, 1993; Kooi and Beaumont, 1994; 1996; Braun and Sambridge, 1997).

$$q = -k\nabla h \tag{1}$$

where k is the mass diffusivity coefficient, expressed in units of area per time [e.g., m^2/y]. Assuming conservation of matter along a 2-D section and no tectonic deformation, h must obey:

$$dh/dt = -\nabla q \tag{2}$$

With constant k, Eqs. (1) and (2) lead to the linear diffusion equation:

$$dh/dt = k\nabla^2 h \tag{3}$$

This model of surface processes holds only for particular conditions. The regolith must form more rapidly than it is removed by surface transport and slopes must not exceed the frictional angle of the material. Even for scarps formed in loose alluvium some complications arise when high scarps are considered. Scarps with height typically in excess of about 10 meters in arid climatic zones, tend to have systematically sharper curvatures at crest than at base (e.g., Andrews and Bucknam, 1987). Gravity-driven erosion processes such as hillslope landsliding impose strong limitations on the applicability of the diffusion equation since the processes are rather slope- then curvature-dependent, which basically requires to introduce slope-and-height dependent terms in the equation (3). At the larger scale, hillslope and stream processes interact and the sediment transport then depends nonlinearly on the slope and on other factors such as the slope gradient, the area drained above the point, the distance from the water divide, so that the simple 2-D linear diffusion does not apply in general (e.g., Gossman, 1976). In spite of these limitations, we have chosen to stick to a linear diffusion law to model erosion in the upland. This model does not accurately mimic the spatial distribution of denudation in the mountain range but it leads to a sediment yield at the mountain front that is roughly proportional to the mean elevation of the basin relative to that point (a rough approximation to the sediment yield resulting from a change of elevation h over a horizontal distance d is $k \times h/d$) and therefore accounts for the apparent correlation between elevation and denudation rates (Ahnert; 1970, Pinet and Souriau, 1988; Summerfield and Hulton, 1994). We did not apply the diffusion model to the whole system, however. We felt that we should take into account the major discontinuity in surface processes that occurs at the mountain front. As a river emerges into the adjacent basin its gradient is sharply reduced and deposition occurs. The streams shift from side to side and build up alluvial fans and tend to form a broad gently sloping pediment at the base of the mountain range. In addition, a lateral drainage often develops along the foothills of mountain ranges. The Ganges along the Himalayan foothills, the Parana along the Andes, or the Tarim along the Tien Shan are good examples. Altogether the formation of the pediment and lateral drainage tend to maintain gentle slopes in the foreland. There is therefore a sharp contrast between river incision that maintains a rugged topography with steep slopes in the mountain range and widespread deposition of alluvium in the foreland. This discontinuity of processes must be considered to model the sharp break-in-slope at the mountain front that is generally observed on topographic profiles across mountain belts. In order to simulate this major change in surface processes, sedimentation in the lowland is modelled assuming flat deposition by fluvial network: we assume that conservation of matter along the section and the sediment at the mountain front is distributed in order to maintain a flat horizontal topography in the foreland. We arbitrarily set the change from diffusional erosion to sedimentation ("flat deposition") at a differential elevation of 500 m, which is, however, representative for the transition from highlands to forelands.

We considered values for k varying between 10^2 to 10^4 m^2/y that yield denudation rates of the order of a few 0.01 mm/y to 1 mm/y for a 200–400 km-wide range with a few thousand meters of relief. In order to test the sensitivity of our model on the assumed erosion law we also considered non linear erosion laws of the form:

$$dh/dt = k^*(x,h,\nabla h)\nabla^2 h \tag{4a}$$

where $k^*(x,h,\nabla h) = k(x)(\nabla h)^n$ (e.g., Gossman, 1976; Andrews and Bucknam, 1987). We will refer to the cases with n = 1, 2 as first-and second-order diffusion, respectively. In these cases we did not introduce the change in regime at the mountain front since the nonlinear effects already tend to form relatively smooth pediments. It should be noted that Eq. (4) differs from the one obtained assuming a non linear diffusion coefficient in Eq. (1). In that case conservation of mass would lead to an additional term $\nabla k^* \nabla h$:

$$dh/dt = k^*(x,h,\nabla h)\ \nabla^2 h + \nabla k^*(x,h,\nabla h)\nabla h \tag{4b}$$

However, Eq. 4a is a phemenological one and may reflect the possibility of material loss from the system. It is also noteworthy that the existing nonlinear erosion laws are not limited to Eq. 4a (e.g., Newman, 1983; Newman et al., 1990), which only presents the simplest way to account for dependence of erodibility on the morphology.

Long-range surface processes. The long-range surface processes are associated with fluvial transport, i.e., with river incision, slope geometry, character of sediment matter, and conditions for deposition (Flint, 1973; 1974; Sheperd and Schumm, 1974; Hirano, 1975; Schumm et al., 1987; Seidl and Dietrich, 1992; Govers 1992a,b; Hairsine and Rose, 1992; Sklar and Dietrich, 1998;2001; Howard et al., 1994; Howard, 1998; Smith, 1998; Davy and Crave, 2000; Snyder et al., 2000; Snyder, 2001; Hancock and Willgoose, 2001; Simpson, 2004). The characteristic laws for this range are different as these mechanisms are dependent on the incision and transport capacity of the fluvial network, local slope, and type of sediment. Deep steep rivers can carry sediment longer distances as it can be caught in turbulent flow layer. Shallow rivers would deposit sediment rapidly resulting in rapid river blockage and frequent change of the direction of the fluvial network. There is also a strong dependence of transport capacity on the grain size and climate episodicity (e.g., Davy and Crave, 2000). The long-range fluvial models were used with success by Kooi and Beamount (1994; 1996), Garcia-Castellanos (2002), Garcia-Castellanos et al. (2002; 2003), Persson et al. (2004). The cumulative material flow, q_{fe}, due to the fluvial transport can be presented, in most simple form, as:

$$q_{\mathrm{fe}} = -K_r\, q_r\, dh/dl \qquad (4c)$$

where q_r is the river discharge, K_r is nondimensional transport coefficient and dh/dl is the slope in the direction of the river drainage with l being the distance along the transporting channel. The diffusion equation (4a), except if it is not strongly nonlinear, provides symmetrical, basically over-smoothed shapes whereas the fluvial transport equation (4c) may result in realistic asymmetric behaviors, because, locally, the direction of each bifurcation of the fluvial network is affected by negligibly small factors, even though the overall direction of the flow is controlled by the regional slope of topography (Fig. 3). Any important change in the regional slope of topography, such as at the transition from tectonically built steep highlands to flat sedimentary built forelands, may result, at some moment, in a drastic change of the direction of the fluvial network, which may choose a principally new stream direction orthogonal to the highland network (as it is the case for the Ganges river, for example). This happens when the sedimentary basin is filled to a point that the inclination of its surface in the direction of tectonic convergence becomes less important than that in some other direction (basically in the direction of the boundary between the steep highlands and flat lowlands).

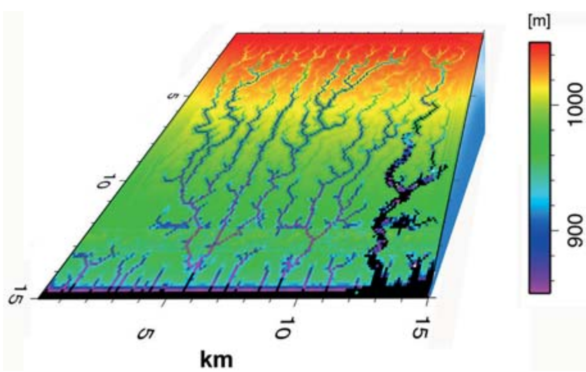


Fig. 3. Example of a typical numerical morphology model with surface erosion and sedimentation based on linear diffusion erosion equation and fluvial transport equation (Poisson, et al., 1996). Diffusion equation, except if it is not strongly non-linear, provides symmetrical shapes whereas fluvial transport equation may result in asymmetric behaviour because, locally, the direction of each bi-furcation of the fluvial network may be affected by negligibly small factors, even though the overall direction of the flow is controlled by the regional slope of topography

Although river networks in mountain ranges owe their existence to the competing effects of tectonic uplift and climate-controlled erosion, it was also argued that some universal geometric properties of river networks may be relatively independent of both tectonics and climate (Casteltort and Simpson, 2006). These authors have proposed that the geometry of river networks is established on the lowland margins of incipient uplifts, and is quenched into the erosion zone as the mountain belts widen with time. In that model, the geometry of river networks simply reflects the downward coalescence of alluvial rivers on undissected surfaces outside of mountain belts, and is therefore independent of erosion processes. Yet, the amount of the transported matter, incision rates, and other major dynamic parameters of the network are definitely tectonic-and-climate dependent.

3.2 Alternative Models of Surface Processes

The diffusion equation reflects an integrated effect of various processes acting at micro-and macroscale: chemical and physical erosion and weathering, gravity hillslope sliding etc. Some of these processes, for example, chemical erosion, are well described by the diffusion equation, since it reflects the physics of propagation of chemical interactions. On the other hand, gravity-driven processes are not diffusive. These processes are primarily slope dependent and thus do not fit well within the linear diffusion model (Densmore et al.,1997; 1998; Hasbargen and Paola, 2000; Roering et al., 2001; Schorghofer and Rothman, 2002; Pelletier, 2004). Indeed, it has been noted that the diffusion equation tends to over-smooth the predicted topography and fails to reproduce the usually sharp transitions from tectonically modified uplifted landscape to typically flat deposition surfaces in the foreland basins. To remedy this problem, either enhanced split (bi-mode) erosion models that discriminate between diffusion and gravity driven processes (e.g., Simpson and Schlunegger, 2003) or alternative stochastic (based on methods of artificial intellect such as cellular automates) and analogue models were proposed (Crave et al., 2000; Davy and Crave, 2000; Crave and Davy, 2001; Bonnet and Crave, 2003). Crave et al. (2001) or Tucker and Bras (1998; 2000), for example, used stochastic methods based on cellular automats that "learn" how to reproduce erosion/sedimentation from pre-imposed logical rules that establish relations between a given grid cell and its neighbors, as a function of the local slope, height, precipitation, regolith type and other conditions. If the rules and their relations are well established, they may form the "vocabulary" and "grammar" (= "language") for description of topography evolution. This approach may eventually produce more realistic landscapes than the common diffusion-fluvial transport models. However, for each new application, it requires one to justify the local applicability of the previously established rules. Analogue (physical) erosion models were used to study erosional response to tectonic forcing (e.g., Lague et al., 2003). These models may produce naturally looking landscapes, yet their applicability is rather limited since it is highly difficult to control, scale and interpret their parameters.

Linear and nonlinear diffusion short-range models combined with fluvial transport long-range models (Fig. 3) remain to be most widely used for tectonic-scale modelling. In particular, diffusion and fluvial transport equations can be generalized (Simpson and Schlunegger, 2003) as following:

$$\begin{cases} \dfrac{dh}{dt} = \nabla \cdot \left((k + cq^{n})\nabla h\right) \\ \nabla \cdot \left(\dfrac{\nabla h}{|\nabla h|}\, q\right) = -\alpha \\ De = \dfrac{c\alpha^{n} L^{n}}{k} \end{cases} \tag{5}$$

where c is sediment discharge, α is effective rainfall, q is surface fluid discharge, and $k + cq^n$ has a sense of a variable nonlinear diffusion coefficient that incorporates both the effects of diffusion-driven (k-term: chemical and physical erosion, weathering) processes and gravity-driven, i.e. fluvial, processes (cq^n term: slope-dependent flow, sliding, creep etc). The respective role of dispersive processes and hillslope creep processes is characterized by dimensionless De number (L is characteristic length scale).

4 Structure and Rheology of the Lithosphere

4.1 Rheology

Many studies of the interplay between erosion and tectonics have been conducted assuming either local isostasy (Ahnert, 1970; Leeder, 1991) or thin plate flexural behavior of the lithosphere (Beaumont, 1981; Flemings and Jordan, 1989;1990; Beaumont et al., 1992; Masek et al., 1994a,b; Garcia-Castellanos, 2002; Garcia-Castellanos et al., 2002; Garcia-Castellanos et al., 2003). Some authors have considered the possibility for ductile flow in the lower crust and treated the lower crust as an inviscid fluid overlaid by a thin elastic plate (King et al., 1988; King and Ellis, 1990; Avouac and Burov, 1996; Burov and Cloetingh, 1997; Burov et al., 2001). The effect of variations in the surface loading and in the crustal thickness on the mechanical behavior of the lithosphere have been often neglected, except several studies (e.g., Beaumont et al., 1992, 2000; Avouac and Burov, 1996; Burov and Cloetingh, 1997; Burov et al., 2001; Toussaint et al., 2004a,b). The coupled erosion-tectonics regime described in the previous sections assumes that strain localization below a tectonic load, range or basin, results from weakening of the lithosphere due to crustal thickening and bending stresses. In order to account for this process one can treat the lithosphere neither as a one-layer elastic or visco-elastic plate with vertically integrated properties overlying an inviscid asthenosphere, or as a thin viscous sheet (e.g., England and McKenzie, 1983; Vilotte et al., 1982). We thus have to consider the lithological and mechanical rheological layering of the lithosphere. For the model demonstrated here, three lithological layers were defined: the upper crust, the lower crust, and the mantle (Fig. 2a). Each layer has specific properties (density, mechanical, and thermal constants) that are given in Table 1. We assume no compositional changes due to deformation or cooling. The lithological boundary between the upper and lower crust lies at a fixed depth of 20 km. The bottom of the mantle lithosphere is limited by the 1330°C isotherm at a depth of about 250 km. At small differential stresses the rocks behave elastically. In terms of principal components, the relationship between the stress tensor, σ, and the strain tensor, ε, can be written:

$$\sigma_j = 2\mu_e \varepsilon_j + \lambda(\varepsilon_1 + \varepsilon_2 + \varepsilon_3) \tag{6}$$

where j =1, 2, 3. λ and μ_e are Lamé's constants related to Young's modulus (E) and Poissons's ratio ν as $\lambda = E\,\nu((1+\nu)(1-2\nu))^{-1}$; $\mu_e = E/2(1+\nu)$. Typical values for E and ν are 6.5–8×10^{10} N/m^2 and 0.25, respectively (e.g., Turcotte and Schubert, 1982).

Weakening by brittle failure or ductile flow occurs when elastic stresses reach some threshold value that determines the condition for failure or significant ductile deformation. Above this threshold rocks no longer behave elastically, and unrecoverable strain may grow without increase of stress. The conditions of brittle failure are independent of rock type and temperature, but strongly controlled by pressure (Byerlee, 1978):

$$\begin{aligned} &\sigma_3 = (\sigma_1 - \sigma_3)/3.9 \text{ at } \sigma_3 < 120\text{ MPa}; \\ &\sigma_3 = (\sigma_1 - \sigma_3)/2.1 - 100\text{ MPa at } \sigma_3 \geq 120\text{ MPa} \end{aligned} \tag{7}$$

where σ_1, σ_2, σ_3 are principal stresses [MPa]. This law corresponds to Mohr-Coulomb plastic behavior.

Ductile flow in the lithosphere essentially results from dislocation creep (e.g., Kusznir, 1991). This

Table 1a. Definition of variables

Variable	Values and units	Definition	Comments
$\tau_{xx}, \tau_{xy}, \tau_{yy},$	Pa, MPa	shear stress components	
$\sigma_{xx}, \sigma_{xy}, \sigma_{yy}$	Pa, MPa	full stress components	$\boldsymbol{\sigma} = \tau - P\mathbf{I}, \sigma_{xy} = \tau_{xy}$ etc.
P	Pa, MPa	pressure	
$\mathbf{v}$	m/s, mm/y	total velocity vector	
u	m/s, mm/y	horizontal velocity	x component of v
v	m/s, mm/y	vertical velocity	y component of v
μ	Pa s	effective viscosity	10^{19} to 10^{25} Pa s
k	m^2 / y	coefficient of erosion	~ mass diffusivity
dh	m, km	topographic uplift	or subsidence
du	m, km	tectonic uplift	do not mix with u
de	m, km	erosion	or sedimentation
ψ	m^2/s	stream function	$\upsilon = \partial\psi / \partial y, v = \partial\psi / \partial y$
ξ	s^{-1}	vorticity function	$\partial\upsilon / \delta y - \partial v / \partial x = \Delta\psi$
ε		strain	
$\dot{\varepsilon}$	s^{-1}	average strain rate	$\dot{\varepsilon} = (½\, \dot{\varepsilon}_{ij}\dot{\varepsilon}_{ij})^{½}$
q	m^2/s	integrated flux	ductile crust
q_e	(m^2/s) / m	erosional flux	per unit length
E	8×10^{10} N/m^2	Young's modulus	in the semi-analytical model
v	0.25	Poisson's ratio	in the semi-analytical model
λ, μ_e	N/m^2	Lamé's constants	
A^*	Pa^{-n} s^{-1}	material constant	power law
n	3 to 5	stress exponent	power law
H^*	kJ mol^{-1}	activation enthalpy	power law
R	8.314 J/mol K	gas constant	power law
T	C°, K	temperature	
$\gamma(y)$	Pa/m, MPa/km	depth gradient of yield stress	$\gamma(y) \propto d\sigma(\varepsilon)/dy,$
w	m, km	plate deflection	~ deflection of mantle lithosphere
$T_e, \tilde{T}_e\,(x, w, w', w'', t)$	m, km	effective elastic thickness	~ instant integrated strength
T_{ec}	m, km	effective elastic thickness of the crust	$T_e \approx (T_{ec}^3 + T_{em}^3)^{1/3}$ $T_{ec} \le h_{c_1}$
T_{em}	m, km	effective elastic thickness of mantle lithosphere	$T_e \approx (T_{ec}^3 + T_{em}^3)^{1/3}$ $T_{em} \le h_{c_2} - T_c$
$\tilde{M}_x$	N m / m	flexural moment	per unit length
$\tilde{T}_x$	N	longitudinal force	
$\tilde{Q}_x$	N / m	shearing force	per unit length
p^+	Pa, N/m^2	surface load	
p^-	Pa / m	restoring stress	per unit length
$h\,(x, t)$	m, km	surface topography	
$\tilde{h}\,(x, t)$	m, km	upper boundary of ductile channel	
h_c, T_c	m, km	Moho depth	Moho boundary
h_{c_2}	m, km	lower boundary of ductile crustal cannel	$h_{c_2} \le T_c$