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Syn-convergence flow inside and at the margin of orogenic plateaus: Lithospheric-scale experimental approach

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Abstract. This study investigates three-dimensional flow modes of orogenic plateaus by means of physical modeling. Experiments consist of shortening two contiguous lithospheres of contrasting strength, one being a weak plateau-type lithosphere, and the other a strong craton-type lithosphere. The lateral boundaries are either totally confined or allow escape toward a lateral foreland on one side. Two syn-convergence flow regimes are distinguished, which are governed by the balance between the gravity potential and the strength of the plateau crust and the resistance of its lateral foreland. The first regime implies transversal (orogen-normal) injection of plateau lower crust into the collision zone as a result of confinement of the plateau by an increasingly stiffer lateral boundary. As a precursor mechanism to channel flow, transversal injection responds to downward thickening of the plateau crust that is forcibly extruded into the orogenic wedge. The second regime is that of collapse-driven lateral escape of the plateau. This regime is established after a threshold is attained in the inter-plate coupling in the collision zone, which allows the gravity potential of the plateau to overcome the resistance of its lateral boundary. Under the collapse-driven escape regime (orogen-parallel), such as that governing Tibet during the last 13 Ma, most of the convergence in the plateau and the top and rear of the collisional wedge is transformed into lateral flow and extension.

Keywords: crustal flow, analog modeling, lateral escape, extrusion tectonics, orogenic plateau, hot orogen.

1. Introduction

Having attained a plateau stage, orogens are hot, owing to their high content in heat-producing elements due to crustal thickening [England and Thompson, 1984] and/or
because their mantle lithosphere has been thinned by thermal erosion or removed by
delamination [England and Houseman, 1989; Molnar et al., 1993; Sandiford and
Powell, 1991; Figure 1]. Orogenic plateaus are thus particularly weak due to a reduced
temperature-dependent viscosity of the crust and/or to regional partial melting
[Vanderhaeghe and Teyssier, 2001]. As weak lithospheric regions submitted to
convergence such as long-lived accretionary orogens, cordilleras and wide mature
collision zones, orogenic plateaus deform dominantly by flow, which tends to maintain
a low-relief surface and Moho by responding coevally to body forces due to the
topography and surface forces due to convergence [e.g., England and Houseman, 1989;
Bird, 1991; Chardon et al., 2009].

Lateral flow in orogenic plateaus combines orogen normal shortening, forced
tectonic extrusion (driven by forces applied at boundaries) and collapse-driven escape
(gravitational flow) in a context of partial or full mechanical coupling between the
upper mantle and the upper crust [Royden, 1996; Holt, 2000; Tikoff et al., 2004;
Andronicos et al., 2007; Chardon et al., 2011]. Concomitantly, orogen-normal extrusion
of plateau lower crust into the adjoining collision wedge is inferred to occur along a
channel bounded by a thrust at its base and a detachment at its top [i.e., the channel flow
hypothesis; Grujic et al., 1996; Beaumont et al., 2001]. One must therefore consider the
fundamentally three-dimensional aspects of both lateral flow inside the plateau and its
potential interplay with transverse flow at the edge of the plateau. Investigating three-
dimensional deformation of hot orogens is permitted in physical experiments where a
thin/hot model lithosphere is shortened between two quasi-rigid walls [vice models;
Cruden et al., 2006; Cagnard et al., 2006a; Riller et al., 2012]. Similar experiments
have been performed with a cold/thick lithosphere [Davy and Cobbold, 1988].
Nonetheless, a limitation of vice models is that they simulate the shortening of a single
type of lithosphere and do not reproduce collision of a plateau with a colder, deformable
lithosphere that would produce an orogenic wedge at the plateau edge. To test three-
dimensional deformation modes of an orogenic plateau, the introduction of such
rheological contrasts should be convenient to simulate mechanical coupling in the
collision zone, which is critical to understand strain partitioning between the plateau and
its foreland [Dewey, 1988] and potential transverse mass transfers across the plateau
edge [Beaumont et al., 2004]. Ratschabacher et al. [1991a] modeled lateral flow in
experiments where a strong foreland and a non-deformable indenter flank a weak
lithosphere, with variable degree of lateral confinement. However, the foreland remains
undeformed in almost all experiments and the study focuses on lateral escape.

The present study investigates syn-convergence deformation and flow modes
operating inside and at the margin of orogenic plateaus or hot orogens using lithosphere
scaled-analog models. The experiments involve shortening of two contiguous
lithospheres of contrasted strengths, a weak one and thin plateau-type lithosphere, and a
strong, thick and yet deformable craton-type lithosphere. Our experimental setup is
designed to investigate internal deformation of a weak plateau lithosphere being
shortened under various degrees of lateral confinement. It also permits deformation of
the plateau to interact with the stiff lithosphere in the collision zone by simulating
various degrees of coupling between the two lithospheres.

We focus on the three-dimensional aspects of mass transfers within and at the
collisional margin of the model orogenic plateau. We show that the evolving strength of
the lateral foreland of the plateau primarily governs the relative magnitude of
transversal (orogen-normal) and longitudinal orogenic flow. We also describe a
potential precursor mechanism to channel flow that consists of forced transversal
extrusion of the plateau lower crust as a consequence of downward directed tectonic thickening of the plateau upper crust.

2. Experimental setup and scaling

2.1. Model design and experimental procedure

The model lithospheres float on sodium polytungstate mixed with glycol water, which is a dense, Newtonian low-viscosity material modeling the asthenosphere (Figure 2). Ductile lithospheric layers are modeled with silicone putties (GSIR gums from Rhone-Poulenc, France) that are Newtonian fluids at experimental strain rates [Weijermars and Schmeling, 1986]. Dry Fontainebleau sand mixed with ethyl-cellulose powder is used to simulate the brittle upper crust [Davy and Cobbold, 1991; Table 2].

The strong cold cratonic lithosphere consists of a brittle upper crust, a ductile lower crust and a thick and dense high-viscosity lithospheric mantle (Figure 2a). The latter simulates the upper part of the lithospheric mantle, which concentrates most of the strength. The weak hot plateau lithosphere is modeled as a three-layer crust (i.e., a lithosphere lacking its mantle part). The model consists of a brittle upper crust, a ductile mid-crust and a low-viscosity lower crust simulating partial melting-induced weakening (Figure 2a). The model is placed in a Plexiglas box equipped with a moving-wall coupled with a piston and a motor to control shortening. Velocity discontinuities made of Plexiglas plates placed on each lateral wall of the box help to localize deformation away from the moving-wall and to limit boundary effects (Figure 2). Eleven experiments were performed in laterally confined conditions (Figure 2a, Table 1). Three experiments were performed at a medium to low lateral confinement, allowing lateral escape of the model during convergence (Figure 2b). In that case, the same type of model as that of laterally confined experiments was placed in a larger box, adjacent to a
confining silicone capable of absorbing the escape of the shortened lithospheres (Figure 2b). All experiments were performed at constant room temperature. There is no thermal evolution of the system, density contrasts between the layers remain constant, and we neglect the role of thermal diffusion and phase changes.

During each experiment, photographs were taken in order to map successive stages of surface deformation (e.g., Figure 3). At the end of the experiment, the model is freezed for at least 12 hours. There is no significant post-experiment deformation because: (1) shortening is no longer active, (2) the viscosity of the silicone decreases rapidly with decreasing temperature and the sudden drop in temperature prevents it from flowing, (3) comparison between the last top view photo and the frozen cross-sections shows similar faults locations. Then the lithospheric part of the model is detached from the sodium polytungstate, buried in sand and wetted. The protective sand-shell is also frozen, the block formed by the model lithosphere and the sand is then cut with a circular saw to obtain serial cross-sections. For practical reasons related to model preparation, the two lithospheres share the ductile crustal layer of intermediate viscosity (DC; Figure 2). In the cratonic lithosphere, this layer models the whole lower crust, whereas in the weak lithosphere it simulates the middle crust. Model setting implies that the weak lithosphere is initially weaker than the craton and bears a 50 km-thick crust at the onset of shortening. Such a setting is convenient to model an advanced stage in collision and crustal thickening, which is for instance consistent with the Neogene configuration of the Tibet-Himalaya system [e.g., Replumaz et al., 2010c].

2.2. Scaling, limitations of the models and tested parameters

In order to perform experiments comparable with nature, geometric and dynamic similarities must be respected [Davy and Cobbold, 1991]. This is verified if the relation
\[ \sigma^* = \rho^* g^* L^* \]  
(1)

is respected [Brun, 2002], where \( \sigma^*, \rho^*, g^*, L^* \), are the ratio between model and nature for stress, density, gravity and lengths, respectively. All experiments were performed under natural gravity, thus \( g^* = 1 \). The scaling factor for length is \( L^* = 5 \times 10^{-7} \), i.e. 1 cm in experiments is 20 km in nature. In the brittle layers, the differential stress obeys the Coulomb criterion. In the ductile layers, the shear stress is

\[ \tau = \eta \frac{d\varepsilon}{dt} \]  
(2)

where \( \tau \) is the shear stress, \( \eta \) the shear viscosity, and \( \frac{d\varepsilon}{dt} \) the shear strain rate.

The shear strain rate \( \frac{d\varepsilon}{dt} \) depends strain localization (faulted or undeformed regions) and evolves through time. A good approximation of the bulk shear strain rate (for the entire model and whole duration of the experiment) is

\[ \frac{d\varepsilon}{dt} \sim \frac{V}{L} \]  
(3)

where \( V \) is the imposed shortening rate and \( L \) the length of the shortened area. Equation (2) is then equivalent to

\[ \tau \sim \eta \frac{V}{L} \]  
(4)

In all experiments, the brittle layer is modeled using a mix of dry sand and ethyl-cellulose powder whose cohesion is negligible. According to Schellart [2000] and Lohrman et al. [2003], the friction coefficient of sprinkled sand varies between 0.5 and 1.2 for sifted sand and for the range of normal stresses considered in our experiments. In order to estimate the friction coefficient of the sand we evaluated the dip of reverse and normal faults in simple experiments, and we concluded that the friction angle is \( \sim 30^\circ \) for the sprinkled sand we used, i.e. the friction coefficient is close to 0.6. The viscosity of the silicone layer that models the ductile lithosphere varies between \( 9.5 \times 10^3 \) Pa s and \( 3.81 \times 10^4 \) Pa s (Table 2). Figure 2 shows the strength profiles of the weak and cratonic lithospheres for the reference experiment, strength values being summarized in
Table 3. The overall model strength profiles for the lithospheres respect the scaling law (Figure 2) and makes our experiments comparable with nature. However, the ratio between the viscosity of lithospheric layers and that of the asthenosphere is $\sim 3 \times 10^4$, which is more than one order of magnitude larger than in nature [Loiselet et al., 2009; Schellart, 2010]. In our experiments, the asthenosphere essentially exerts normal hydrostatic stresses on the lithosphere. This should not affect fundamentally the evolution of our experiments, since neither a subduction zone nor any large deformation appear at the lithosphere/asthenosphere boundary in the models. In our experiments, lithospheric deformations only result from the applied boundary conditions and from the collapse of lithospheric plates, not from asthenospheric fluxes.

Model lithospheres respect the scaled ratio of crust density over lithospheric mantle density (Table 2). For practical reasons, the density of the asthenosphere in the models is higher than that of the lithospheric mantle (1.75 and 1.56 respectively) whilst they are of the same order in nature. This has an impact on the gravity potential of the lithospheres that is overestimated. We calculated Argand numbers ($Ar$) characterizing the ratio of body forces over tectonic forces, to check if body forces alone are large enough to result in lateral escape (Appendix A, Table 4).

$$Ar = \frac{F_B}{F_T}$$

If $Ar > 1$, the lithosphere is unstable and can potentially collapse; whereas for $Ar < 1$ the body forces may be supported by the strength of the lithosphere. This number is increased by a factor ~2 when the density of the liquid modeling the asthenosphere rises from 1.56 to 1.75. Hence, lateral escape is favored in our experiments if permitted by lateral boundary conditions (unconstrained experiments). In nature, many other phenomena that are not reproduced in experiments may also favor lateral escape, as for instance, the subduction of an oceanic plate of the lateral foreland that would modify the
lateral traction exerted on the collision zone, as in SE Asia for the Tibetan plateau. Our unconfined experiments reproduce the deformation of a continental collision zone in which the lateral tectonic boundary conditions favor lateral escape. On the other hand, confined experiments show the evolution of a craton/plateau collision zone in which lateral boundary conditions oppose it. Another limitation of our models is that they do not reproduce the early stages of oceanic and continental subduction, which implies that we neither take into account the influence of inherited structures from these stages, nor simulate processes such as pre-collisional subduction, back arc extension, or slab-break-off. We choose analog materials that reproduce a mature plateau (weak and thick crust), corresponding to the collisional stage. Although no oceanic subduction occurs, the continental cratonic lithosphere underthrusts the plateau along a large-scale shear zone and reproduces a realistic collision dynamics at the suture (see below).

The impact of several parameters has been tested: (1) the shortening velocity, which sets the strength of the ductile layers and the degree of brittle-ductile lithospheric coupling, (2) the thickness and viscosity of the crustal ductile layers to test different strength contrasts amongst crustal layers and between the plateau and cratonic lithospheres, and (3) the thickness of the silicone layer that opposes lateral escape (Table 2).

3. Results

Performing an experiment with the same experimental settings twice and stopping them at two different amounts of shortening allowed us to reconstitute successive deformation stages in cross-section view. Comparison of the serial cross-sections of a same model also helps constraining the sequential development of the structures given the lateral gradient of finite shortening existing between the sides of the box (near the
velocity discontinuity generated by the tip of the lateral wall of the piston) and its center
(Figures 2 and 3). In all the figures, the weak lithosphere is placed on the left hand side
of the maps and cross-sections. In the text, pro-thrusts designate thrusts verging towards
the cratonic foreland (the pro-side of the orogen), whilst retro-thrusts designate thrusts
verging towards the hinterland (the retro-side of the orogen).

3. 1. Laterally confined experiments

3. 1. 1. Structures and kinematic evolution

There is a remarkable consistency in the development mode of the structures and
flow patterns in all the laterally confined experiments. Upper crustal shortening first
affects the weak lithosphere, which absorbs ca. 70% of the total shortening. From 3-4%
of shortening, a first series of retro-thrusts forms in the weak lithosphere, each fault
evolving into a conjugate couple of reverse faults delimiting pop-downs [e.g., Figure 4c
and 4d; see Cagnard et al., 2006b]. Each conjugate fault set is abandoned when a new
pop-down forms. In experiments with contrasted brittle crust thicknesses between the
two types of lithospheres (Figures 4b to 4d, 4f and 5), the last conjugate fault set forms
at the lithosphere boundary at 10-15% shortening. It remains active until the end of the
experiment. The retro-thrust of this late conjugate couple is predominant.

Shortening of the cratonic crust is localized along a limited number of thrusts
delimiting pop-ups, which become generally inclined or overturned cratonwards 3-5%
of shortening after they form (e.g., Figures 3, 4a, 5). Ductile layers are generally
thickened, especially the low-viscosity lower crust in the weak lithosphere, whose
thickness is doubled after 30% of shortening (Figures 3 and 4). The mid-crust of the
plateau is thickened as much as its weak lower crust; whereas the ductile cratonic crust
undergoes less thickening (25% after 30% of bulk shortening).
The plateau lower crustal layer is thrust upon the cratonic lithospheric mantle, initiating continental scale underthrusting of the cratonic mantle along a pro-shear zone. This lithospheric shear zone is shown in Figures 3 to 5 by white arrows. The plateau lower crustal layer keeps a flat floor and thickens as a symmetrical dome (Figure 3) that becomes overturned above the lithospheric pro-shear zone (Figures 4a to 4d). Asymmetrical doming results in a wedge shape of the plateau lower crust, whose roof makes the back limb of the dome and dips toward the hinterland. As the dome amplifies, it is obliquely injected into the cratonic lower crust, which is in turn underthrust beneath the inverted limb of the dome (Figures 4a to 4d). A far-field channel flow effect is produced in the cratonic lower crust by injection of plateau lower crust into the foreland (e.g., Figure 4e).

The back-limb of the dome shows upright mullions or folds interpreted as a consequence of distributed shortening (Figures 4a to 4d). This attests to limited crustal-layering parallel shearing at the roof of the weak lower crust and to dominant homogeneous thickening of the plateau at the rear of the dome. This also means that part of the mid crust of the weak lithosphere is transported cratonward on the roof of the weak lowermost crust. The top of the dome, once amplified, is taken into a flat retro-shear affecting the upper mid-crust, which is marked by backward deflection of former mullions or folds at the dome’s crest (Figures 4b, 4d, 5). This shear zone emerges as the last active retro-thrust making the boundary of the plateau crust (Figures 4b, 4d, 5). The boundary between the two types of brittle crusts migrates twice as much more towards the hinterland than the intersection between the underthrusting shear plane and the asthenosphere. This explains the large magnitude of mid- and upper crustal back shearing above the overflowing lower crust of the weak lithosphere. In the cratonic lithosphere, the last pro-thrust propagates upward from the underthrusting shear zone.
Importantly, no net unroofing of the plateau weak lower crust occurs as a consequence of doming and lower crust overflow. Indeed, the crest of the dome does not attain shallower depths than the initial depth of the top of the plateau weak lower crustal layer.

3. 1. 2. Sensitivity to parameters

The shortening velocity, by controlling the degree of coupling between the lithospheric layers, affects the style of deformation. For a high velocity, thrusts remain active for a shorter time period and are abandoned more quickly. An increase in coupling (higher velocity) is also expressed by earlier faulting in the cratonic crust, and a greater dip and lower finite displacement of the lithospheric shear zone (45° for 4 cm h\(^{-1}\) in Figure 4c against 20° for 0.4 cm h\(^{-1}\) in Figure 4d). An increased strength (i.e. velocity) also leads to less backward displacement of the suture. Therefore, for a high degree of coupling, shortening is preferentially accommodated by thickening instead of by underthrusting and/or injection.

Using the same thickness of the brittle crust over the entire model favors pro-thrusts and pop-downs between 20 and 30% of shortening (Figures 3 and 4a). Finally, changing the thickness and rheology of the plateau crust to highly over-thickened and less viscous (Figures 4e and 4f) favors distributed deformation and limited stacking of pop-downs in the plateau. In this situation, instead of dipping towards the hinterland, the roof of the plateau lower crust is either flat or cratonward dipping, which suggests an active role of the gravitational potential of the thickened crust on the dynamics of cratonward extrusion.

3. 2. Experiments with lateral escape
3.2.1. Structures and kinematic evolution in map view

In this section, we will first describe experiment F2 (Figure 6), taken as a reference experiment. We will then evaluate the effect of an increased shortening velocity (experiment F3, Figure 7) and of an increased strength of the silicone layer opposed to lateral escape (experiment F1, Figure 8). The term external refers herewith to the part of the model located close to the boundary where lateral escape is permitted; whereas the term internal, which refers to the confined boundary side of the model.

From the beginning of the experiment, the entire model escapes laterally, forming an outwardly convex arch. Deformation nucleates as conjugate strike-slip fault sets affecting both lithospheres (Figure 6). The main intersection between these faults defines a sag basin marking the transition between an internal domain dominated by reverse faulting and transpression, and an external domain dominated by extension and transtension (Figure 6). As shortening increases (16%), the plateau lithosphere concentrates the deformation by developing an anastomosing network of longitudinal transpressive faults (pro and retro) that propagate towards the external domain in the form of conjugate strike-slip faults (Figure 6). Between 7.5 and 21 % shortening, a retro-thrust nucleates and propagates at the boundary of the two lithospheres. The hinterland-side fault system of the plateau is abandoned and the rest of the plateau fault network is shortened and laterally elongated as a result of fault rotations. A new dextral-oblique thrust propagates obliquely across the cratonic lithosphere by joining the piston and the plateau. This fault defines the front of the external triangle-shaped half of the cratonic lithosphere that is deformed. Continuous widening and clockwise rotation of the grabens attest to combined lateral stretching and dextral shearing of that triangle which responds to faster lateral stretching and escape of the plateau lithosphere compared to the cratonic lithosphere. Overall shortening induces partitioning of the
deformation between the two lithospheres. The cratonic lithosphere undergoes limited 327 shortening and heterogeneous lateral extension. The weak lithosphere deforms by bulk 328 homogeneous pure shear deformation, which is accommodated by an anastomosing 329 network of transpressive faults [Cagnard et al., 2006a; Cruden et al., 2006; Riller et al., 330 2012]. To summarize, transverse shortening and lateral escape of the plateau by bulk 331 pure shear induce a combination of heterogeneous simple shear and pure shear in the 332 cratonic lithosphere (Figure 9).

Increasing the strength of the ductile layers by increasing the velocity of shortening 333 leads to a lesser amount of lateral escape (experiment F3; Figure 7), especially if 334 compared to the amount of shortening (Figure 10d). The smaller thickening/escape ratio 335 in this experiment results from the weaker influence of body forces due to density 336 contrasts compared to forces arising from convergence velocity. The plateau lithosphere 337 undergoes more escape than the cratonic lithosphere. The retro-thrust at the lithospheric 338 boundary does not propagate beyond the central sag basin (Figure 7). Deformation 339 concentrates along the hinterland side of the plateau and over its entire length by 340 activation of successive straight thrusts (Figure 7). On both sides of the sag basin, two 341 large areas of the plateau remain nearly undeformed against the hinterland deformation 342 belt (Figure 7). Though not visible, deformation occurs at the lithospheric contact in the 343 external part of the model since diffuse shear is required along this boundary in order to 344 explain higher differential escape of the plateau. To summarize, an increased strength of 345 the ductile layers caused by a higher shortening velocity induces (1) smaller amount of 346 lateral escape compared to the amount of shortening, (2) differential escape of the 347 plateau and therefore partial decoupling between the two lithospheres, (3) strain 348 partitioning inside the plateau, and (4) limited deformation of the cratonic lithosphere 349 (Figure 9).
For a strength and thickness of the lateral silicone three times that of experiments F2 and F3, the cratonic lithosphere remains virtually undeformed until late stages of shortening, and the plateau is indented and partly escaped (experiment F1; Figure 8). Bulk pure shear strain of the plateau lithosphere is achieved by an anastomosing network of transpressive faults due to the densification and rotation of an early strike-slip conjugate fault network. Spreading in the escaped plateau peninsula is accommodated by normal faults having a strike-slip component (Figure 8). To summarize, an increase in strength at the lateral boundary favors decoupling between the two lithospheres, pure shear assisted escape of the plateau, and spreading confined to the escaped part of the weak lithosphere (Figure 9).

3.2.2. Structures and kinematic evolution in cross-section

As a significant part of the convergence is absorbed by lateral escape at the earliest stages of shortening, cross-sections of the free boundary experiments provide insights into the earliest stages of convergent deformation. In the inner part of the model, the earliest increments of shortening are accommodated by indentation of the plateau lower crust by the cratonic lithospheric mantle (Figures 6 and 7). The future overflowing dome of plateau lower crust initiates as a result of this indentation, which is also accompanied by doming of the cratonic mantle (Figures 6 and 7). The retro-thrust marking the lithosphere boundary in the upper crust may possibly be related to the early indentation process (Figures 6 and 7). In the external part of the model, the two domes involved in the indentation process either never appeared, or collapsed. The net throw of the retro-thrust is greater than in the internal part of the model (Figures 6 and 7). This is interpreted as a consequence of escape-enhanced lateral thinning of the plateau lithosphere (see below).
3. 2. 3. Three-dimensional analysis

We performed a three-dimensional finite strain analyses (Appendix B) to quantify the deformation along the principal axis of the strain ellipsoid $\lambda_1$ (maximum elongation), $\lambda_2$ (intermediate) and $\lambda_3$ (maximum shortening). It shows that $\lambda_1$ is horizontal and orogen-parallel and $\lambda_3$ is horizontal and orogen-normal in the plateau lithosphere in all experiments with lateral escape (Figure 9). A general transition is documented from vertical flattening in the internal part of the plateau to horizontal plane strain or constriction in the external part of the plateau (Figure 9). For strong coupling between the two lithospheres (experiment F2; Figures 6 and 9), flow toward the lateral boundary compensates thickening, and the external domain of the plateau is deformed by horizontal plane strain. For partial decoupling between the two lithospheres resulting from a higher shortening velocity (experiment F3; Figures 7 and 9), lateral constriction in the external part of the plateau lithosphere attests to a component of thinning superimposed on orogen-normal shortening. Horizontal constriction prevails in the cratonic lithosphere in response to combined lateral stretching and vertical thinning [Merle and Gapais, 1997; Teyssier and Tikoff, 1999].

No cross-sections are available for the indentation experiment characterized by a higher strength of the confining silicone (experiment F1; Figures 8 and 9c), but surface deformation shows that the cratonic lithosphere remains virtually undeformed. Conversely, one may expect stronger vertical flattening in the internal part of the plateau lithosphere. The presence of normal faults in the escaped peninsula (Figure 8, 9c), attests to significant thinning due to spreading. To summarize, lateral finite thinning of the weak lithosphere characterizes the three experiments.
Strain intensities (Appendix B) are of the same order of magnitude for F2 and F3 experiments, and always lower for the cratonic lithosphere (from 0.7 to 1.5) than for the weak lithosphere (from 1.1 to 1.7). In the weak lithosphere, the strain intensity parameter $\varepsilon_s$ [Nadai, 1950] is minimum at the center of the model (Table 4), maximum at the external boundary, and intermediate in the internal domain. This is interpreted as the superimposition of thickening near the confined wall onto a strain gradient due to the laterally increasing component of longitudinal stretching. Higher strain intensity near the lateral boundary is interpreted as the result of the superimposition of a gravitational component of flow onto the bulk shortening strain ellipsoid. Strain intensities in the cratonic lithosphere, though very low, also increase towards the lateral boundary (Table 4).

Lateral escape patterns for strong coupling (F2 experiment) or partial coupling (F3 experiment) between the two lithospheres are comparable (Figures 10b and 10c). Lateral escape is fast at the beginning and decreases rapidly within 5 hours to a nearly steady value (Figures 10b and 10c). In experiment F2, early escape is accommodated by overall thinning of model lithospheres (total escaped area greater than the area lost by the advance of the piston up to 6 hours time; Figure 10b). At the end of both experiments, total lateral escape had roughly accommodated half of the area lost by the advance of the piston, the remaining half of which having been absorbed by lithospheric thickening (Figures 10b and 10c). In the experiment with a thicker confining silicone (F1; high strength of confining silicone, complete decoupling between the two lithospheres, indentation of the weak lithosphere by the craton), lateral escape evolves linearly with time. The escaped surface consists exclusively of weak lithosphere and remains low even at the end of the experiment (50 cm$^2$ vs. 150 cm$^2$ for experiments F2 and F3; Figure 10a to 10c). This analysis suggests that lateral escape is far more
sensitive to the resistance at the lateral boundary than to the strength of the lithospheres, as observed in experiments by Ratschbacher et al. [1991a].

The evolution of the gravity potential and stability of the model lithospheres may be quantified using the Argand number (Ar; eq. 4 and Appendix A). We calculated Ar of the cratonic and weak lithospheres for the initial and final configurations of the experiments (Table 4; Figure 9). This parameter depends on the buoyancy of the lithospheric mantle that is overestimated in our models (Table 2). Consequently, Ar is also overestimated (up to a factor 2), and collapse of the model resulting in lateral escape is favored by the density layering we adopted. As discussed in section 2.2, other processes not reproduced in our models can favor lateral spreading of the plateau. Then, this choice in the parameters may simulate natural situations, for example when lateral oceanic subduction zone occurs while the collision is active, as it is the case in SE Asia.

In all experiments, the cratonic lithosphere is also potentially unstable at the beginning of the experiment (Ar>1). This is attested by the development of grabens and deformation leading to lateral constriction. Nonetheless, the significant differential finite strain and lateral escape undergone by the two lithospheres is enhanced by a higher strength of the cratonic lithosphere compared to the weak lithosphere. Although the cratonic lithosphere is potentially unstable, its resistance prevents a rapid collapse. Collapse efficiency and velocity also depend on the resistance at the lateral boundary and on the degree of coupling between the two lithospheres, which are not taken into account in the expression of the Argand number. Furthermore, when the two lithospheres are coupled (experiments F2 and F3), lateral stretching of the cratonic lithosphere primarily adjusts to lateral escape of the deforming weak lithosphere (Figure 9). The above considerations suggest that despite the experimental limitations, the
dynamic analysis of deformation inside and at the collisional boundary of the weak
lithosphere remains valid.

In experiment F1 including a thick lateral confining layer (Figure 8, Table 4), the
initial negative Argand number of the weak lithosphere is not realistic for simulating an
orogenic plateau. These initial experimental conditions may rather simulate a thin hot
back arc domain precursor of an orogenic plateau. During experiment, the thickening of
the weak lithosphere increases its Argand number, which should in turn favor lateral
escape. Then, the weak lithosphere should not escape right away, but only after a
significant amount of shortening-induced thickening. However, lateral escape takes
place with a constant rate during the entire the experiment (Figure 10a). This apparent
inconsistency shows that horizontal stresses resulting from the shortening of the weak
lithosphere are larger than horizontal stresses arising from density contrasts with the
lateral confining medium. In these conditions, part of the shortening is accommodated
by lateral escape. Such a configuration reflects a forced tectonic extrusion mechanism
such as that modeled by Tapponnier et al. [1982] and in the strongly confined
experiments of Ratschbacher et al. [1991].

For strong (F2 experiment) or partial (F3 experiment) coupling between the two
lithospheres (case of a weaker lateral foreland), the initial Argand number of the weak
lithosphere is significantly higher than that of the cratonic lithosphere (Table 4). At the
beginning of experiments, the main driving mechanism of escape is collapse, as attested
by the evolution of the escape versus thickening ratio through time (Figure 10d).
Collapse-driven escape is fast at the early stages of the experiment. As the lateral
foreland thickens with time, the Argand number of the plateau drops and the escape rate
decreases accordingly and eventually stabilizes (Figure 10b, 10c, and 10d). At that
stage, the models become governed by forced tectonic extrusion in a configuration that
is close to that of experiment F1 with a high strength lateral foreland. In other words, the evolving strength of the lateral foreland exerts a prime control on the dynamics of hot orogens, particularly on their capacity to collapse laterally. Such an influence is similar to that observed by Davy and Cobbold [1988].

4. Discussion

Models describing the dynamics of the Himalaya-Tibet orogen should be consistent with geological observations, earthquakes distribution and deep structure. Several models have addressed these issues. For example Grujic [1996] or Beaumont et al. [2001; 2004] focused on the transverse mass transfers in the collision zone, and Royden et al. [1997, 2008], Clark and Royden [2000] try to explain the topographic gradient and lateral escape at the eastern edge of the plateau. The differences between the frontal and lateral margins arise from boundary conditions (collision with strong lithosphere vs. partially free boundary, respectively), deep structure (underthrusting of India or not), orientation with respect to the direction of convergence (parallel vs. perpendicular) and surface factors (climate, erosion rates). This results in different crustal flow modalities and timing. For the south margin of the plateau, the channel flow model explains exhumation of high grade rocks (High Himalayan Crystalline) bounded by a thrust at its base (Main Central Thrust, MCT) and a detachment at its top (South Tibetan detachment, STD) alongside with focused erosion at the surface. Cratonward injection of plateau lower crust in the collision zone as observed in our 2D models provides insights into the channel flow process (Figures 3 and 4). At the eastern margin of the Tibetan plateau, eastward flow of weak lower crust material without exhumation could explain the topographic gradient and GPS motions in the area. The lateral flow patterns of our 3D models (Figures 6 to 8) allow addressing the mechanism of eastern growth /
escape of the Tibetan plateau. It is important to consider that despite their obvious kinematic differences, both the transverse and lateral flow processes are ultimately driven by gravitational collapse and/or forced tectonic extrusion of plateau lower crust [Bird, 1991; Royden et al., 1997; Shen et al., 2001; Beaumont et al., 2004]. In the discussion, we shall first focus on strain partitioning at the surface and at depth, then on the transversal flow at the India-Asia plate boundary, and finally discuss three-dimensional aspects of the interplay between transversal and lateral flows. As our models do not involve erosion or temperature variations, we shall not discuss surface processes or metamorphism issues.

4.1. Strain partitioning and kinematics

In the Himalaya-Tibet orogen, strain is spatially partitioned, with dip-slip reverse faulting at the orogenic front and transpression in the syntaxes and north of the Kunlun fault, whereas the plateau undergoes transtension, and strike-slip deformation dominates East of the Longmen Shan and Xianshuihe fault [Andronicos et al., 2007]. Our models reproduce well most of this pattern. In 2D and 3D, thrusting initiates at the lithospheres boundary and at the backstop at early stages, with a strike-slip / transpressive deformation component for 3D models forming anastomosed fault networks (e.g. Figures 3 and 6). This confirms recent studies challenging the view of reverse thrusting propagating from the suture to the North [e.g. Tapponnier et al., 2001] and attesting distributed deformation in the plateau with Paleocene/Eocene faulting in north Tibet [e.g. Zhang et al., 2004; Clark et al., 2010]. The weak plateau lithosphere squeezed against the strong Tarim basin / craton in the north, represented by the back wall of the box in our experiments, can deform early far from the collision zone. The Himalayan plate boundary is complex with transpressive deformation along the Karakorum fault,
the suture and the Jiali fault, which is connected with north-south striking grabens from the west – central part of the plateau (Fig). Right-lateral strike-slip shearing (early to mid-Miocene) and east-west extension (mid-Miocene) are very close in age and may have been coeval in east and central Tibet [Murphy et al., 2002, 2010; Taylor et al., 2003; Searle et al., 2011 and references therein]. Our experiments reproduce very well this pattern and are consistent with previous experiments [Ratschbacher et al., 1991a] with the development of a large-scale transpressive fault at the boundary between craton and plateau lithospheres, coeval with the opening of conjugate grabens (Figures 6 and 7). The grabens rotate clockwise to accommodate lateral escape but stay roughly parallel to each other and do not show changes in strike from west to east as in Tibet [Kapp and Guynn, 2004]. The plate boundary stays almost straight unlike the arcuate Himalaya bounded by syntaxes. This implies that processes not reproduced in the experiments and controlled by earlier oceanic subduction stages and/or the shape of Greater India [Ali and Atchinson, 2005; Replumaz et al., 2010c; van Hinsbergen et al., 2011; Bajolet et al., 2013] would be responsible specific features of the orogen such as (i) the Himalayan orocline and the variable strike of rifts potentially due to oroclinal bending [Ratschbacher et al., 1994], (ii) the lateral variations in strain orientation along the orogenic front [Seeber and Pecher, 1998; Kapp and Guynn, 2004] and (iii) southward propagation of the orogenic front [Murphy and Copeland, 2005; Bajolet et al., 2013]. In our models, the grabens are also restricted to the plate boundary area and within the cratonic lithosphere, and transpression is confined to the inner plateau (Figures 6 to 8), whereas central Tibet is under transtension [Andronicos et al., 2007]. The boundary between these domains is also more irregular in nature than in our experiments [e.g. Ratschbacher et al., 2011]. Possibly because in our experiments the rheological contrast between the plateau and the craton localizes this boundary.
Moreover, the Tibetan plateau's width is variable (from ca. 400 km near the west syntaxe to 1000 km in its center) whereas it is constant in our experiments. In our 3D models, gravitational collapse is mainly accommodated by lower ductile flow, whilst it is also partially accommodated in the upper crust in Tibet by active grabens. Transtension is observed in the less deformed external part of the 3D models that rotates clockwise during escape (Figures 6 to 8). Experiments F2 and F3 (Figure 6 and 7) with a thin confining layer show large amount of escape as in east Tibet but limited rotation. Experiment F1 with a thick confining layer (Figure 8) reproduces well the overall rotation and deformation pattern of extruded SE Tibet between the Red River and Sagaing faults, but with a small amount of escape. This would confirm the heterogeneity in lithospheric strength east of Tibet, with an overall weak lateral foreland allowing eastward escape, but bearing a strong Sichuan basin resisting and redirecting lateral escape to the SE [Cook and Royden, 2008; Robert et al., 2010a].

If it is now well established that deformation is distributed within the Tibetan plateau, strain localization at depth is still debated, especially in eastern Tibet. Earthquakes distribution [Clark and Royden, 2000; Steck et al., 2001; Andronicos et al., 2007; Priestley et al., 2007] and geophysical imaging [Bai et al., 2010; Liu et al., 2014; Bao et al., 2015] suggests that 1) the mid/lower crust is heterogenous with weak regions (interconnected or not) and stronger ones, 2) some faults are rooted in the deep crust and channelize crustal flow, whereas others are restricted to the upper crust. In our experiments, the fact that the most external (escaped) part the models remains mostly undeformed and that silicone layers are homogeneous does not allow us to model complex flow patterns at depth and test the hypotheses mentioned above.

4.2. Transverse mass transfers at plateau edge
Our experimental results help reconstruct a sequence of deformation at the contact between a hot and a cold lithosphere under ongoing convergence such as the Himalaya collision zone (Figure 11). Early shortening is absorbed dominantly by indentation of the hot lithosphere by the mantle layer of the cold lithosphere. In the weak lithosphere, ongoing convergence is accommodated by distributed thickening and by burial of upper crust by the formation of pop-downs [Cagnard et al., 2006b]. Pop-downs are not observed in the cratonic lithosphere because strong lithospheric mantle prevents downward motion of crustal material. Instead, shortening is accommodated by thickening and pop-ups. Indentation at depth evolves into the subduction of the cratonic mantle beneath the weak lithosphere. Decoupling along the continental subduction enables overflow and later injection of the lower crust of the hot lithosphere into the orogenic wedge. Shortening of the cold lithosphere induces moderate crustal thickening and strain localization along foreland-verging thrusts rooted in the subduction shear zone. The shear zone accommodating back-shearing at the roof of the overflowing hot lower crust emerges as a retro-thrust at the suture. In other words, transverse advection of weak lithosphere lower crust produces indentation of the foreland crust by coeval activation of (1) retro-fold-and-thrust belt whose front coincides with the suture, and (2) a pro- (subduction) shear zone rooting in the crust-mantle boundary. The surface trace of the suture zone migrates towards the hinterland concomitantly with cratonward injection of lower crust and continental subduction. Interestingly, these features and kinematics are also produced in channel flow numerical experiments where the craton lower crust does not subduct with, and is decoupled from, the underlying lithospheric mantle and the trench is fixed [Beaumont et al., 2004]. In numerical models with trench advance as in the India-Asia collision, the motion of the suture (proward or retroward) depends on the trade-off between the subduction velocity and the denudation intensity.
at the orogenic front. Our experiments do not simulate early stages of subduction and
the cratonic upper crust is not forced to subduct beneath the plateau lithosphere as in
numerical models, which may explain the migration of the surface trace of the suture
toward the hinterland. However, the suture, branching on a lithospheric shear zone,
controls underthrusting of the cratonic lithospheric mantle at a shallow angle and a fold-
and-thrust belt develops at the lithospheres boundary. This configuration is similar to
the India-Asia collision where Indian mantle underthrusts the Tibetan lithosphere over a
large area [Jimenez-Munt et al., 2008; Li et al., 2008]. One may infer from the
experiments that in case of efficient erosion at the orogenic front, the dome of lower
plateau crust could be significantly exhumed. Such an exhumation would stop backward
flow at the roof of the dome and may change the kinematics at the suture. Therefore, the
trench-fixed dynamics reproduced in the experiments may be only transient and coeval
with dome amplification and ascent, before trench-advance occurs in late collisional
stages.

Keeping a nearly stationary flat Moho and topographic surface, the plateau crust
shortens and thickens by preferential downward motions that are transformed into
cratonward (i.e., transverse) extrusion of lower crust (Figure 11). The initiation and
amplification mode, structure and kinematics of the lower crustal dome indenting the
collisional wedge compare to those of hot fold nappes and may reflect early stages of a
channel flow process. However, in contrast with fold nappes that are fed from below by
upward extrusion of collisional wedge crust [Merle and Guillier, 1989; Duretz et al.,
2011], our model implies feeding of the transversally extruding lower crust by
downward burial of thickened plateau crust.

In channel flow models as set up by Beaumont et al. [2004], the shape and amplitude
of cratonward flow are determined by the conditions at the plate boundary and the
denudation intensity at the orogenic front. Exhumation of lower crust is achieved for
denudation rates of ca. 1 cm yr\(^{-1}\) and a fixed shape of the subduction fault assuming a
non-deformable upper mantle. These parameters ensure the formation of a large-scale
channel in the mid-crust (injecting the craton over several hundreds of kilometers)
unlike the structure observed in our models. The fact that no net exhumation of the
dome takes place in our models requires either a change in the boundary conditions of
the injection, or an additional phenomenon not modeled here and permitting the hot
lower crust to reach the surface. An increase of mechanical coupling between the two
lithospheres will inhibit subduction and transversal flow. This should also favor the
nucleation of a thrust ramp below the advancing dome, which would form a fault bent
fold involving the dome, thereby enhancing focused erosion and ultimately exhumation
of the dome (Figure 11). No structure comparable to the South Tibetan Detachment
(STD) is formed during our experiments, most probably because the exhumation of the
dome is not complete. However, toppling of the backlimb of the dome on the ramp
would turn the retro-thrust system accompanying transversal injection into a hinterland-
dipping detachment system comparable to the STD (Figure 11, e.g., Godin et al., 2011).
Whatever the mechanism controlling the kinematics of this detachment (gravitational
collapse or extension-assisted mountain building), the formation of a crustal ramp and
its consequences could set favorable conditions to establish channel flow as envisaged
by Beaumont et al. [2001, 2004]. In the scenario exposed above, back thrusts at the rear
of the exhumed lower crust (former dome) may either be fossil (abandoned after
activation of the detachment) or reactivated by gravity-driven slip along the detachment.
Reactivation of the Himalayan suture along back-thrusts of post-early Oligocene age
[Mascle et al., 1986; Yin et al., 1994; Aldsdorf et al., 1998] may indicate the emergence
of the retro fold-and-thrust belt produced experimentally.
4. 3. Three-dimensional mass balance: transversal vs. longitudinal crustal flow

The results of the two sets of experiments suggest that three-dimensional mass redistribution takes place by combining transversal injection of plateau lower crust into the frontal foreland as modeled by Beaumont et al. [2001], and lateral flow of the weak lithosphere as proposed by Royden et al. [1997] (Figure 11). The relative magnitude of the two processes is controlled by a balance between the degree of mechanical coupling along the craton/plateau boundary and by lateral forces resulting from the resistance and buoyancy of the lateral foreland. We observe that lateral escape observed in our 3D experiments prevents significant continental underthrusting and concomitant injection of plateau lower crust into the orogenic wedge. This implies that transversal (injection/subduction) and longitudinal (escape) mass transfers may not coexist with comparable fluxes during the entire evolution of the orogenic system. For an unstable lithosphere, gravity forces can be accommodated by faulting in the upper crust and/or by ductile flow at depth. Stages of transversal-dominated and escape-dominated flow may alternate depending on the boundary conditions at the orogenic front and lateral boundary. In the following, we propose a typical scenario for the three dimensional evolution of a collision zone between a cratonic and a weak lithosphere.

The strength of the lateral foreland and forces arising from density contrasts on the lateral boundary of the converging system primarily control the mode of lateral escape of the weak lithosphere. As in our experiments, during the early stages of collision, a weak lateral foreland and a large Argand number of the hot lithosphere enhance lateral collapse-driven escape even for low values of shortening-induced thickening (configuration of F2 experiment). A lateral flow regime is installed. In nature, lateral escape could also be favored by lateral tension forces arising for instance from
subduction occurring close to the lateral boundary of the collision zone [e.g.,
Tapponnier et al., 1986; Ratschbacher et al., 1991b; Fournier et al., 2004; Schellart
and Lister, 2005; Faccenna et al., 2006]. Strength built-up in the lateral foreland as a
consequence of escape-driven tectonic thickening may reach a threshold, triggering
thickening of the weak lithosphere and slow forced tectonic extrusion (F1 experiment
configuration). At this stage, decoupling in the collision zone favors a transversal flow
regime driven by transformation of plateau overthickening into cratonward expulsion of
lower crust (Figures 4, 5 and 11).

The transversal flow regime is maintained as long as the stiff lateral foreland limits
collapse. During this period, gravity forces are accommodated by cratonward injection
of plateau lower crust between a thrust and a detachment, as the High Himalayan
Crystalline, whose exhumation was driven by coeval and opposed shearing along the
MCT and STD between ca. 25 and 13 Ma. In details, the exhumation process was
probably more complex, not fully synchronous along the range, and likely of lesser
amplitude than that simulated in Beaumont et al. [2004] and in our experiments,
especially considering the lateral structural / rheological heterogeneity of the lower crust
[Long and McQuarrie, 2010; Rey et al., 2010; Kellett and Grujic, 2012].
A return to a lateral flow regime would be permitted if a change in the configuration
of the orogenic system would again permit collapse-driven lateral escape. Such a
change may result form a threshold in the gravitational potential of the weak lithosphere
with respect to its lateral foreland and/or from a drop in the degree of coupling at the
craton/plateau boundary. Such a process may have resulted from break-off of the Indian
subducting slab inferred to propagate from west to east between 25 and 10 Ma
[Replumaz et al., 2010b; Stearns et al., 2015]. Slab break-off may largely increase the
elevation of the plateau [e.g., Molnar et al., 1993], its Argand number compared to its
lateral foreland and, in turn, favor lateral escape. The lateral flow regime would be installed at the end of the break-off or once the process is completed. Initiation of a subduction at the lateral boundary could also enhance lateral escape.

A reorganization of the frontal continental subduction fault system may also cause for switching from a transversal to a lateral orogenic flow regime. Nucleation and activation of a crustal ramp in the footwall of the transversally overflowing lower crust (see above) could trigger an increase in the inter-plate coupling, an increase in topography, and therefore favor collapse-driven lateral escape. Abandonment of the MCT around 13 Ma and activation of the Main Boundary Thrust (MBT) could have triggered such a change.

The switch from a transversal to a lateral orogenic flow regime is exemplified in the Ama Drime Massif of the High Himalayan Crystalline, which has been exhumed at the southern edge of the Tibetan plateau between the MCT and the STD. Exhumation of the massif started at ca. 30 Ma along the STD with N-S extension, and the tectonic regime then switched to E-W extension activating the NS-striking normal shear zones bounding the massif at ca. 12-13 Ma [Jessup et al., 2008; Cottle et al., 2009; Kali et al., 2010]. The age of ~ 13 Ma is consistent with the onset of lateral extension in the Tibetan plateau estimated between 16 and 8 Ma [Armijo et al., 1986; Styron et al., 2015] and attributed to lateral collapse [e.g., Coleman and Hodges, 1995; Williams et al., 2001]. Lateral collapse therefore postdates the main episode of transversal flow of the High Himalaya Crystalline between 25 and 15 Ma [Burg et al., 1984; Hubbard and Harrison, 1989; Hodges et al., 1992]. The formation of the MBT in the footwall of the MCT around 13 – 12 Ma [e.g., Kali et al., 2010] would have provided the necessary conditions for a change from a transversal to longitudinal flow regime in the Himalaya-Tibet orogenic system. This change may have been accompanied by a transient flow
regime during which the extruding lower crust of the High Himalaya Crystalline
recorded both transversal flow at depth and lateral flow at its roof [Burg et al., 1984;
Brun et al., 1985; Gapais et al., 1992; Beaumont et al., 2001; Murphy and Copeland,
2005]. The onset of normal faulting is also documented at ca. 13 Ma in other Himalayan
domes [e.g. Murphy et al., 2002; Languille et al., 2014] and in Tibet [e.g. Armijo et al.,
1986; Blisniuk et al., 2001; Styron et al., 2015]. Rifts and extensional focal mechanisms
are more numerous in south than in central or north Tibet [Copley et al., 2011]. We
propose that after 13 Ma, gravitational collapse was mainly accommodated by brittle
faulting in the upper crust in south Tibet and by ductile lateral flow in northern Tibet.
This can be understood considering that 1) lateral flow in south Tibet is partially
blocked by the east Himalayan syntaxis, 2) southern Tibet is underthrusted by India,
herewith increasing its overall strength. In our experiments, grabens are indeed more
developed in the plateau lithosphere for overall stronger ductile layers (Figures 6 and 7).
This would be consistent with recent studies highlighting the importance of India
underthrusting on rifting in Tibet [Copley et al., 2011; Ratschbacher et al., 2011; Styron
et al., 2015]. If orogen-parallel extension is partly accommodated by brittle faulting in
northern Tibet, the main mechanism is ductile lateral flow along strike-slip faults, as in
our 3D experiments (Figure 6 to 8). In areas where the lithosphere has regained a stable
thickness, lateral escape could continue, driven by forced tectonic extrusion.

5. Conclusions

The present work suggests that paired orogenic plateaus – collisional wedges may
undergo two distinct syn-convergence flow regimes governed by the balance between
the gravity potential and the strength of the plateau crust, and the resistance of its lateral
foreland. A first possible regime, accretion dominated, is characterized by transversal
injection of plateau lower crust into the collisional wedge as a result of confinement of
the plateau by an increasingly stiffer lateral foreland. In this regime, transversal
injection is driven by downward thickening of the plateau crust that is forcibly extruded
into the orogenic wedge. The second orogenic flow regime is characterized by collapse-
driven lateral escape of the plateau as a consequence of increasing inter-plate coupling
in the collision zone, which increases the gravity potential of the plateau with respect to
the resistance of its lateral foreland. During collapse-driven lateral escape, a large
proportion of convergence-induced thickening of the plateau and the top of the
collisional wedge is transformed into lateral constrictional flow and extension.

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Appendix A. Calculation of the Argand number

The Argand number allows the quantification of the stability of the lithosphere. It is expressed as the ratio between body forces that favor the gravitational collapse of the lithosphere, and tectonic forces that oppose it \cite{England1982, Houseman1986}:

\[ \text{Ar} = \frac{F_B}{F_T} \]  \hspace{1cm} (1)
For our models, the buoyancy force may be calculated as the difference between the gravitational potential energy of the lithosphere (cratonic or plateau type) integrated over depth and that of the lateral foreland (confining silicone):

\[ F_B = \int_{\text{surface}}^{z_{\text{surface}}} (\rho(z) z g)_{\text{collision}} \, dz - \int_{\text{surface}}^{z_{\text{surface}}} (\rho(z) z g)_{\text{lateral}} \, dz \]  

Given the collapse regime operating in the models with lateral escape, the tectonic force is calculated assuming a vertical \( \sigma_1 \) as the sum of tectonics forces necessary to stretch the brittle and ductile layers:

\[ F_T = F_{T_{\text{brittle}}} + F_{T_{\text{ductile}}} \]  

with

\[ F_{T_{\text{brittle}}} = \frac{1}{3} g \rho_{\text{BC}} h_{\text{BC}}^2 \]  

\[ F_{T_{\text{ductile}}} = 2 \eta h \frac{d\varepsilon}{dt} \]  

where \( g \) is the gravity, \( \rho_{\text{BC}} \) and \( h_{\text{BC}} \) the density and thickness if the sand layer, \( \eta \) and \( h \) the viscosity and thickness of the silicone layer, and \( d\varepsilon/dt \) the experimental strain rate.

The strain rate \( d\varepsilon/dt \) is calculated as

\[ d\varepsilon/dt = V/W \]  

where \( V \) is the shortening velocity and \( W \) the length of the model in the shortening-parallel direction. Thicknesses of layers at the end of each experiment were measured on the most internal and most external cross-sections; whereas those at the beginning of the experiments are known from model building. This allows us to quantify the stability of the lithospheres at the beginning and at the end of the experiments. Argand numbers are given in Table 4.

Appendix B. Quantification of deformation
Following the method proposed by Cruden et al. (2006), a three-dimensional finite strain analysis has been performed by measuring deformation of the ductile layers on each cross-section of the models. For this purpose, the x direction is defined as the direction of convergence, y as the direction of lateral escape (orogen parallel), z as the vertical direction. Initial and final measured lengths (L₀ and L) and thicknesses (T₀ and T) of the ductile layers give the principal strains in x and z directions with the relation

\[ \lambda_x = \left(\frac{L}{L_0}\right)^2 = (1+\varepsilon_x)^2 \quad \text{and} \quad \lambda_z = \left(\frac{T}{T_0}\right)^2 = (1+\varepsilon_z)^2 \]  

(1)

The principal strain in the direction of lateral escape (y) is calculated from the change in cross-sectional area can be expressed as

\[ \lambda_y = \left(\frac{A}{A_0}\right)^2 = (1+\varepsilon_y)^2 \]  

(2)

where A₀ and A are the initial and final area respectively. Flinn’s K parameter defining the aspect ratio of the strain ellipsoids has been calculated as

\[ K = \frac{\lambda_1}{\lambda_2-1}\left/\frac{\lambda_2}{\lambda_3-1}\right. \]  

(3)

\(\lambda_1, \lambda_2\) and \(\lambda_3\) being the maximal, intermediate and minimum principal axes of the strain ellipsoid, respectively. Strain intensity for each ellipsoid is estimated by \(\varepsilon_s\) [Nadai, 1950], defined as

\[ \varepsilon_s = \sqrt{3}/2 \times \gamma_0 \]  

(4)

with

\[ \gamma_0 = 2/3 \times \left\{ [\ln(\lambda_1/\lambda_2)]^2 + [\ln(\lambda_2/\lambda_3)]^2 + [\ln(\lambda_3/\lambda_1)]^2 \right\}^{1/2} \]  

(5)

Note: ¹. See supplementary material comprising photos, descriptions of the experiments that were not detailed in the paper and tables containing strain and lateral escape measurement data.
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Table captions

Table 1. Experimental parameters.

Table 2. Material properties for experiments laterally confined and with lateral escape.

Table 3. Differential stress of the ductile lithospheric layers for two different experimental shortening rates.

Table 4. Strain intensities ($\varepsilon_s$) and Argand numbers (Ar) calculated for experiments with lateral escape. Argand numbers are given for initial and final configurations for the most internal and external parts of the models. For experiment F1, lack of cross-sections did not allow us to calculate the final Ar and $\varepsilon_s$. 
Figure captions

Figure 1. a) Map of the main tectonic features of the Tibet-Himalaya orogenic system. Grabens and detachment faults are highlighted in red. F: fault; BNS: Bangong-Nujiang Suture; IT: Indus-Tsangpo; STD: South-Tibetan detachment; MCT: Main Central Thrust; MBT: Main Boundary Thrust; MFT: Main Frontal Thrust; LMSF: Longmenshan fault.

b) Synthetic cross-section. BC: brittle crust; WLC: weak lower crust; CLM: cratonic lithospheric mantle.

Figure 2. Experimental setup. The graphs are strength profiles of both types of lithospheres. BC: brittle crust; DC: ductile crust; WLC: weak lithosphere lower crust; CLM: cratonic lithospheric mantle. Strength values are given in Table 3.

Figure 3. Top views and cross-sections of experiment C1 (see Tables 1 and 2 for parameters). Top views (top) show selected stages of fault surface pattern development, that on the right showing the configuration at the end of the experiment. On cross-sections, the numbers and percentages in bracket refer to the order of appearance and amount of shortening at which each fault formed, respectively. The lines of sections (5, 3) are located on the last top view. WL: weak lithosphere; CL: cratonic lithosphere; d: dome (see text).

Figure 4. Cross-sections of experiments C2, C4, C5, C6, C7 and C8 (see Tables 1 and 2 for parameters). White dashed lines represent originally horizontal passive markers in the ductile layers. Ductile lower crust in the weak lithosphere of experiments C7 and C8.
(TWLC in Tables 1 and 2) is initially doubly thickened. In experiment C7 (panel 4e),
the markers slightly lighter than the DC tend to localize the deformation. Moreover, the
model is deformed by post-experiment downwarping. See Figure 3 for caption.

**Figure 5.** Top views and cross-sections of experiment C3 (see Tables 1 and 2 for
parameters). White dashed lines represent passive markers in the ductile layers that
were originally horizontal. See Figure 3 for caption.

**Figure 6.** Top views and cross-sections of experiment F2 (see Tables 1 and 2 for
parameters, and Figure 3 for caption).

**Figure 7.** Top views and cross-sections of experiment F3 (see Tables 1 and 2 for
parameters, and Figure 3 for caption).

**Figure 8.** Top views of experiment F1 (see Tables 1 and 2 for parameters).

**Figure 9.** Summary of results of experiments with lateral escape. See text for further
explanations.

**Figure 10.** Relationships between lateral escape, thickening and thinning through time
for experiments with lateral escape. (a) to (c) Lateral escape of the model lithospheres
through time based on top view analysis. The portion of the area lost by the advancing
piston that is absorbed by lithospheric thickening results from the subtraction of the
total lithospheric area escaped to the area lost by advancing piston. (d) Evolution of the
thickening/escape ratio versus shortening.
Figure 11. Idealized view of two-stage orogenic flow mode at the edge of an orogenic plateau. See text for further explanations.
Laterally confined experiments

Experiments with lateral escape
Initial thickness and position of the model

Cross-section view

- Brittle crust
- Ductile crust (DC)
- Weak lower crust (WLC)
- Cratonic lithospheric mantle (CLM)
- Passive markers in ductile crust
- Neutral silicone
- Thickened weak lower crust (TWLC)
- Thrust fault
- Normal fault
- Strike-slip fault
- Shear sense
- Boundary between WL and CL brittle crusts

Plane view

- Fault
- Thrust fault
- Normal fault
- Indentation

C1 settings:
- laterally confined
- 30% shortening
- 2cm/h
- homogeneous sand thickness

10% shortening
20%
30%

Velocity discontinuity (VD)

Initial thickness and position of the model
17% shortening

C3 settings:
- laterally confined
- 50% shortening
- 0.75 cm/h
F2 settings:
- free lateral boundary
- 1 cm of neutral silicone
- 25% shortening
- 0.75 cm/h
8% shortening

initial width

16%

25%

F3 settings:
- free lateral boundary
- 1 cm of neutral silicone
- 25% shortening
- 2 cm/h

10 cm
F1 settings:

- free lateral boundary
- 3 cm of neutral silicone
- 40% shortening
- 0.75 cm/h
INCREASED OVERALL DUCTILE STRENGTH OF LITHOSPHERES

Ar = 7.1
vertical flattening
\lambda_2 > 1
horizontal plane strain

Ar = 2.7
vertical flattening
\lambda_2 = 1
horizontal plane strain

Ar = 6.3
horizontal plane strain

Ar = 0.5
horizontal constriction

\lambda_2 < 1
vertical flattening

\lambda_2 = 1
horizontal plane strain

\lambda_2 < 1
horizontal constriction

\lambda_2 > 1
\lambda_2 = 1
\lambda_2 < 1

Ar = 7.1
vertical flattening
\lambda_2 > 1
horizontal plane strain

Ar = 2.7
vertical flattening
\lambda_2 = 1
horizontal plane strain

Ar = 6.3
horizontal plane strain

Ar = 0.5
horizontal constriction

\lambda_2 < 1
vertical flattening

\lambda_2 = 1
horizontal plane strain

\lambda_2 < 1
horizontal constriction

\lambda_2 > 1
\lambda_2 = 1
\lambda_2 < 1

Weak lithosphere
upper + ductile crust

Cratonic lithosphere
upper + ductile crust

Cratonic lithosphere
lithospheric mantle

Weak lithosphere
lower crust
<table>
<thead>
<tr>
<th>Table 1.</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Experiment</strong></td>
</tr>
<tr>
<td><strong>Confined boundary</strong></td>
</tr>
<tr>
<td>C1</td>
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<tr>
<td></td>
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<tr>
<td></td>
</tr>
<tr>
<td>C2</td>
</tr>
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<td></td>
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<tr>
<td>C3</td>
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<tr>
<td>C4</td>
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<tr>
<td>C5</td>
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<tr>
<td>C6</td>
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<td>C7</td>
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<td>C8</td>
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<tr>
<td><strong>Lateral escape</strong></td>
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<tr>
<td>F1</td>
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<tr>
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<td></td>
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<tr>
<td></td>
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<tr>
<td>F2</td>
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<tr>
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<tr>
<td></td>
</tr>
<tr>
<td>F3</td>
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</tr>
</tbody>
</table>

* The properties of each silicone are listed in Table 2. CL – cratonic lithosphere; WL – weak lithosphere; DC – ductile crust; WLC – weak lower crust; TWLC – thickened weak lower crust; CLM – cratonic lithospheric mantle; CS – confining silicone.
### Table 2.

<table>
<thead>
<tr>
<th>Materials</th>
<th>Viscosity (Pa s)</th>
<th>Density</th>
<th>Thickness (cm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sand + ethyle cellulose</td>
<td>-</td>
<td>1.37</td>
<td>0.4 or 0.8</td>
</tr>
<tr>
<td>Sodium polythungstate + glycol water</td>
<td>1.2</td>
<td>1.75</td>
<td>7</td>
</tr>
</tbody>
</table>

**Silicones**

<table>
<thead>
<tr>
<th>Materials</th>
<th>Viscosity (Pa s)</th>
<th>Density</th>
<th>Thickness (cm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ductile crust (DC) blue</td>
<td>$3.57 \times 10^4$</td>
<td>1.40</td>
<td>1</td>
</tr>
<tr>
<td>Weak lower crust (WLC) purple</td>
<td>$2.77 \times 10^4$</td>
<td>1.42</td>
<td>1</td>
</tr>
<tr>
<td>Thickened weak lower crust (TWLC) brown</td>
<td>$9.51 \times 10^3$</td>
<td>1.42</td>
<td>2</td>
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<tr>
<td>Cratonic lithospheric mantle (CLM) pink</td>
<td>$3.58 \times 10^4$</td>
<td>1.56</td>
<td>2</td>
</tr>
<tr>
<td>Confining silicone (CS)</td>
<td>$5.8 \times 10^3$</td>
<td>1.48</td>
<td>1 or 3</td>
</tr>
<tr>
<td>Layer</td>
<td>$\sigma_{1-\sigma_3}$ (Pa)*</td>
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<tr>
<td>----------------------------</td>
<td>-------------------------------</td>
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<td></td>
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<tr>
<td></td>
<td>Shortening rate of 2 cm h(^{-1})</td>
<td>Shortening rate of 0.75 cm h(^{-1})</td>
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<tr>
<td>Weak lithosphere</td>
<td></td>
<td></td>
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</tr>
<tr>
<td>Ductile crust</td>
<td>4.97</td>
<td>1.86</td>
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</tr>
<tr>
<td>Lower crust</td>
<td>3.90</td>
<td>1.46</td>
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<tr>
<td>Cratonic lithosphere</td>
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<td></td>
</tr>
<tr>
<td>Ductile crust</td>
<td>4.97</td>
<td>1.86</td>
<td></td>
</tr>
<tr>
<td>Cratonic lithospheric mantle</td>
<td>5.36</td>
<td>2.00</td>
<td></td>
</tr>
</tbody>
</table>

*Strength profiles for a shortening rate of 2 cm h\(^{-1}\) are shown in Figure 2.
<table>
<thead>
<tr>
<th>Experiment</th>
<th>$\varepsilon_i$</th>
<th>Ar initial</th>
<th></th>
<th>Ar final</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Internal domain</td>
<td>External domain</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Weak lithosphere</td>
<td>Cratonic lithosphere</td>
<td>Weak lithosphere</td>
<td>Cratonic lithosphere</td>
</tr>
<tr>
<td>F1</td>
<td>-3.2</td>
<td>1.0</td>
<td></td>
<td></td>
</tr>
<tr>
<td>F2</td>
<td>1.4</td>
<td>1.1</td>
<td>8.3</td>
<td>4.0</td>
</tr>
<tr>
<td>F3</td>
<td>1.1</td>
<td>0.9</td>
<td>7.2</td>
<td>3.8</td>
</tr>
</tbody>
</table>