Evolutionary model of the Himalaya–Tibet system: geopoem based on new modelling, geological and geophysical data

Alexander I. Chemenda a,*, Jean-Pierre Burg b, Maurice Mattauer c

a Géosciences Azur, UMR 6526, Université de Nice-Sophia Antipolis et CNRS, 250 Rue Albert Einstein, Sophia Antipolis, 06560 Valbonne, France
b Geologisches Institut, ETH, Zurich, Switzerland
c Laboratoire de Géophysique, Tectonique et Sedimentologie, UMR 5573, Université Montpellier II, Case 060, Pl. E. Bataillon 34095, Montpellier, Cedex 05, France

Received 10 August 1999; accepted 19 October 1999

Abstract

A two-dimensional thermo-mechanical laboratory modelling of continental subduction was performed. The subducting continental lithosphere includes a strong brittle upper crust, a weak ductile lower crust, and a strong upper mantle. The lithosphere is underlain by a low viscosity asthenosphere. Subduction is produced by a piston (push force) and the pull force from the mantle lithospheric layer, which is denser than the asthenosphere. The lithospheric layers are composed of material whose strength is sensitive to and inversely proportional to temperature. Throughout the experiment the model surface was maintained under relatively low temperature and the model base at higher temperature. The subduction rate satisfied the Péclet criterion. Modelling confirms that the continental crust can be deeply subducted and shows that slab break-off, delamination and tectonic underplating are fundamental events with drastic consequences on the subsequent evolution of the convergent system. Combining these results with previous, purely mechanical modelling, we elaborate a new evolutionary model for the Himalaya–Tibet convergent system. The principal successive stages are: (1) subduction of the Indian continental lithosphere to 200–250 km depth following subduction of the Tethys oceanic lithosphere; (2) failure and rapid buoyancy-driven uplift of the subducted continental crust from ca. 100 km depth to some depth that varies along the mountain belt (20–30 km on average); (3) break-off of the Indian subducted lithospheric mantle with the attached oceanic lithosphere; (4) subduction/underplating of the Indian lithosphere under Asia over a few to several hundred kilometers; (5) delamination, roll-back, and break-off of the Indian lithospheric mantle; (6) failure of the Indian crust in front of the mountain belt (formation of the main central thrust) and underthrusting of the next portion of Indian lithosphere beneath Tibet for a few hundred kilometers. At the beginning of stage (6), the crustal slice corresponding to the Crystalline Himalayas undergoes ‘erosion-activated’ uplift and exhumation. © 2000 Elsevier Science B.V. All rights reserved.

Keywords: plate collision; tectonics; subduction; physical models; Himalayas; Xizang China; exhumation

1. Introduction

The amount of Indian continental lithosphere consumed in the Himalayas is estimated from pa-
leomagnetic data to be about 1000 km \cite{1,2} and may even reach 1500 km \cite{3}. Recent tomography data \cite{4} are consistent with the latter estimates, revealing high velocity zones that can be interpreted as segments of oceanic lithosphere and continental lithospheric mantle subducted down to 1000 and 1700 km, respectively. It is unclear, however, what is the fate of the buoyant continental crust, which is generally believed to be unable to subduct to these depths. INDEPTH seismic data display the Indian crust underthrust under the Nepalese Himalayas to about 80 km depth \cite{5}. Eclogites in the western Himalayas indicate that the Indian crust had already reached a 100 km depth equivalent (25–30 kbar) at the beginning of collision, ca. 50–55 Ma ago \cite{6,7}. Since then, hundreds of kilometers of the Indian crust have been subducted along the Himalayas. The major challenge now is to understand how this subduction has developed and what is the style of the associated lithospheric deformation. This can be done only on a large scale, considering a long-evolving thermo-mechanical system 'layered lithosphere–asthenosphere'. Such a scale is hardly accessible to geological analysis. Geophysics provide meaningful structural information about the deep lithosphere and mantle but only about the present situation. Experimental and/or numerical modelling thus could play an important role in understanding past events. Modellers (e.g. \cite{8–12}) usually investigate simplified models that cannot consider fundamental factors such as multi-phase (non-stationary) subduction of a large amount of continental lithosphere. Initial stages of this process from subduction of the continental margin to a rapid uplift/exhumation of deeply subducted crust have been experimentally modelled by Chemenda et al. \cite{13,14}. These stages scaled to nature should take place over a few to several million years. The evolution of the Himalaya–Tibet collisional system lasted much longer, 50 to 60 Myr during which a number of other important events may have happened. What are these events and what is their mechanism? Considering the time-scale of the Himalayan history, these questions can be addressed by thermo-mechanical modelling that somehow takes into account changes in the mechanical behaviour of the lithosphere during subduction. The necessity of thermo-mechanical modelling drastically complicates the task and dooms any modelling results to ambiguity. We barely know how the mineralogical composition and rheologic structure of the subducting continental plate change during subduction. These changes strongly depend on a number of ill-understood factors such as the kinetics and spatial distribution of mineralogical reactions \cite{15}, changes of the grain size that strongly affects the rheology, the effect on rheology of fluids and the multi-aggregate nature of rocks (e.g. \cite{16}) etc. Therefore, modelling can represent only a qualitative approximation, even if one adopts precise rheologic laws. In this work we attempt simplified thermo-mechanical experimental modelling of continental subduction under different ‘reasonable’ conditions. The obtained scenarios of initial and mature continental subduction tested against the geological information allowed us to propose a first order physically and geologically consistent model for the evolution of the Himalaya–Tibet system.

2. Mechanical essentials of continental subduction

Previous purely mechanical experiments have revealed two principal factors defining the subduction regime: (1) the strength of the subducting continental crust (upper crust, Figs. 1 and 2) the ratio pull/buoyancy force \( \Omega = F_{pl}/F_{b} \) \cite{13,14}.

The pull force \( F_{pl} \) acting on the subducting continental lithosphere is composed of the traction from both the subducted oceanic lithosphere and the lithospheric mantle of the continental lithosphere. The latter is proportional to both the size (thickness and length in two dimensions) of the subducted segment as well as to the density contrast between the continental and the surrounding mantle. The buoyancy force \( F_{b} \) is proportional to both the thickness and subduction depth of the continental crust as well as to the mantle/crust density contrast. For example, an increase in the \( \Omega \) ratio can be achieved by reduction of the crustal thickness. A decreasing \( \Omega \) ratio can be brought about, in particular, by lowering the lithospheric mantle density. However, the major
factor causing strong and rapid drop in $\Omega$ is break-off of the dense subducted slab. After break-off only the mantle layer of the subducting continental lithosphere can provide the pull force, if this layer is denser than the surrounding mantle. Therefore, the largest pull force is exerted at the beginning of continental subduction whilst the earlier subducted oceanic slab is still attached to the continental lithosphere. Initial continental subduction should be characterised also by comparatively low crustal thickness of the continental margin. Thus, the $\Omega$ ratio should be maximal at this stage.

Purely mechanical experiments corresponding to this situation exhibits the following subduction pattern (Fig. 2): The continental crust subducts to a depth equivalent to 200–300 km, following the oceanic subduction. The upper continental crust then fails near the base of the overriding plate (Fig. 2a) and the subducted crustal slice rapidly raises along and within the interplate zone (Fig. 2b). This buoyancy-driven uplift brings material from several tens to more than 100 km depth equivalent to shallower levels. The depth to which the deeply subducted crust is spontaneously uplifted is proportional to the depth of its initial subduction, which in turn is proportional to the strength of the upper crust. Uplift of the crustal slice is followed by break-off of the previously subducted oceanic lithosphere with part of the continental lithospheric mantle (Fig. 2b). Break-off reduces the pull force, which results in 1–2 km equivalent isostatic uplift of the lithosphere in the subduction zone. Another consequence of the break-off in the experiments conducted at constant convergent rate, imposed by the moving piston (Fig. 1), is an increase in horizontal compression of the lithosphere. Accordingly, the subduction regime was termed low-compressional before, and highly compressional after break-off [14]. The geometry and dynamics of subduction after break-off still depend on the $\Omega$ ratio, which is obviously smaller than before break-off. The subduction angle initially is rather steep, 30° to 60° even if the subducting continental lithosphere is denser than the asthenosphere (Fig. 3a). Then, the subducted lithosphere bends and rotates upwards to come against the base of the overriding plate (Fig. 3b). Horizontal subduction/underplating of this lithosphere follows (Fig. 3c). At a stage that depends on the upper crust strength and on friction between the underplating crust and the

---

**Fig. 1.** Scheme of the modelling. 1, overriding plate; 2, upper continental crust with strong strain weakening; 3, ductile, weak lower crust; 4 plastic mantle layer; 5, piston; 6, liquid asthenosphere; 7, bath. $T_s$ and $T_m$ are the temperatures at the model surface and in the asthenosphere, respectively (in previous purely mechanical modelling discussed in this paper $T_s = T_m$). $h_f$ is the lower crust thickness in the frontal part of the subducting plate.

**Fig. 2.** Low-compressional regime of continental subduction (high pull force), Chemenda et al. [14]. Symbols as in Fig. 1.
overriding plate-sole, the upper crust fails in front of the subduction zone (Fig. 3c). In the conducted experiments the maximal underplating distance $L$ was 7–8 cm corresponding to 250–300 km in nature.

The depth of crustal subduction at which buoyancy-driven lithospheric bending and rotation occurs (Fig. 3b) is proportional to $\Omega$. When the $\Omega$ ratio and the strength of the upper continental crust are such that crustal failure occurs before bending of the subducted lithosphere, convergence proceeds as in Fig. 4. Note that failure in Fig. 4b occurs in front of the subduction zone (highly compressional regime). For the crust to fail at depth $\Omega$ should be higher (low-compressional regime, Fig. 2). Application of erosion to the growing relief modifies the deformation as shown in Fig. 4c,d. Break-off of the mantle lithospheric layer (Fig. 4d) results in the horizontal subduction shown in Fig. 3. We now present thermo-mechanical modelling of this process.

3. Set up of thermo-mechanical modelling

The experimental setting (Fig. 1), model materials, techniques and mechanical similarity criteria are those used in the experiments described above. The principal difference is the introduction of thermo-mechanical elements. Along with purely mechanical similarity criteria, we take into account the Péclet criterion:

$$VH/\kappa = \text{const}$$

where $V$ is the subduction rate; $H$ is the thickness of the plate, and $\kappa$ is the thermal diffusivity of the lithosphere that controls the heating (thermal equilibration) rate of the subducted material. The conductive diffusivity of the lithosphere is about $\kappa^o = 10^{-6}$ m$^2$/s [17], but the actual diffusivity of the subducted material (crust in particular) could be higher, due to the advective heat transfer by fluids. The thermal diffusivity of materials used to model lithosphere is $\kappa^m \approx 8 \times 10^{-8}$ m$^2$/s (here and below superscripts ‘m’ and ‘o’ indicate the model and original parameters, respectively). Let us assume that the geological convergence rate is of the order of 1 cm/yr and the crustal thickness is of the order of 10 km. Then, adopting a model crustal thickness $H^m = 1$ cm, we obtain from criterion (1) that the model convergence rate should be $V^m = 8 \times 10^{-5}$ m/s. For higher effective diffusivity of natural lithosphere, the model convergence rate should be slower. In the experiments presented below we used $V \approx 4 \times 10^{-5}$ m/s.

The realistic thermal regime of the subducting plate is not sufficient, however, for thermo-mechanical modelling. Adequate changes in lithospheric rheology and crustal density during subduction are also needed. The effective strength of the crust should considerably reduce while it descends into the asthenosphere. The reduction factor is indeterminate. In the experiments presented here we have assumed that it is about 5. The analogue material strength is very sensitive to temperature, so that five times reduction of the strength is achieved through raising temperature by 2–3°C. On the other hand, the materials density remains virtually constant. The mechanical experiments above were conducted under a homogeneous temperature near 40°C with a rheologic contrast between the lithospheric layers caused by different layer compositions. In thermo-mechanical experiments (Fig. 1) the model is subjected to a linear, stationary temperature gradient with surface temperature of 39°C or 40°C and 42°C under the lithosphere. The temperature slightly increases with depth in the ‘mantle’ and is 0.2°C higher at
the bottom of the bath than at the base of the horizontal lithospheric layer. During subduction, upper lithosphere layers are heated and, therefore, weakening.

4. Results

Twelve experiments have been conducted under different conditions. We present three representative tests (see Table 1 for parameter values).

Experiment 1, Fig. 5: As in the purely mechanical experiment (Fig. 3b), the subducted slab bends upward to the base of the overriding plate (Fig. 5c) with continental subduction becoming horizontal (Fig. 5d). The subducting crust fails in front of the subduction zone, tectonic underplating of the crust then being relayed by crustal accretion and thickening in front of the subduction zone (Fig. 5d–f). Between stages (e) and (f) two competing processes have been observed: separation (delamination) of the mantle layer from the crust at the front of the subducted crust and failure in the hinge zone of the mantle layer. Finally, failure and break-off of the lithospheric mantle occurred (Fig. 5f). To facilitate delamination (reduce the coupling between the upper crust and the mantle), there are at least two possibilities: one is to reduce the strength of the lower crust by using initially weaker or more temperature sensitive material for this layer. Another possibility is to use the same material, but to increase the thickness of the lower crust \( h_f \) (Fig. 1), which has been done for the next experiment.

Experiment 2, Fig. 6 (Table 1): The crust and the lithospheric mantle split apart from the beginning of subduction, the crust wedges along the base of the overriding plate while the dense mantle layer subducts almost vertically at late stage. This layer undergoes considerable down-dip stretching during sinking but deformation does not localise to cause break-off before the slab-tip reaches the bottom of the bath, when convergence is stopped.

These two experiments show that continental subduction is very sensitive to the rheologic structure of the crust. Small variations in lower crust thickness change drastically the bulk behaviour. It seems obvious that tuning this parameter between the two tested values will result in an intermediate scenario; i.e. subduction starting with horizontal tectonic underplating (as in Fig. 5d,e) and followed by delamination and break-off. Other parameters also affect the result. For example, if the
strength of the mantle layer of the underplated/subducted lithosphere in Expt. 1 (Fig. 5) was higher, then one would expect backward delamination and peeling of this layer instead of rupture. Break-off would occur later, when the length of the sinking mantle lithosphere is long enough to generate a sufficient pull force for breaking this strong layer. Such a process is observed in the next experiment where the same lithospheric model as in Expt. 1 is tested under lower temperature and hence with higher strength of the lithospheric layers (Table 1).

Experiment 3, Fig. 7: The initial stages of this experiment are similar to Expt. 1, but the subducting lithospheric mantle separates from the tectonically underplated crust and peels back to the subduction front (Fig. 7f). This delamination is followed by break-off of the mantle layer, which could not be achieved because the sinking mantle layer reached the bottom of the box. The continental crust underplated the overriding plate (Fig. 7f) over 10 cm (~350 km in nature) and did not fail because it is colder and stronger than in Expt. 1.

One can test many other parameter combinations, but the presented experiments provide enough information to predict other possible options for lithospheric/crustal behaviour: the principal elements are delamination, roll-back and break-off of the lithospheric mantle layer. The distance $L$ of tectonic underplating depends on the crustal strength. Weakening of the crust with depth/temperature results in shorter $L$ values. As
we do not know the actual weakening factor we cannot predict the underplating distance in nature. The peel-back distance of the mantle layer is particularly controlled by the lower crust properties. This process can propagate to the subduction zone (Fig. 6) or mid way (Fig. 7) or go farther below the continent (we obtained such a result in experiments not presented in the paper).

To complete the scope of continental subduction scenarios we remind below one more relevant result from purely mechanical modelling [18]. This modelling takes into account the existence of a thin and weak lithosphere in the arc area and shows subduction of most of the fore arc block (Fig. 8a). Evidences for this process have been found in the Urals and the Variscan belt [19], the Kamchatka [20], Taiwan [18,21] and the Himalayas [22,23]. The fore arc block shields from the hot mantle the deeply subducted continental crust (Fig. 8d) which, therefore, is kept relatively cold and hence strong. We did not yet apply thermo-mechanical modelling technique to arc-continent collision and will consider below that the subducting continental crust behaves at this stage as in isothermal experiments (Fig. 2).

5. Evolutionary model of the Himalayas

We use the physically possible subduction scenarios presented above to construct an evolutionary model for the Himalaya–Tibet collisional system (Fig. 9).

The Indian continental margin arrived against Asia 55 to 60 Ma ago [24], resulting in arc-con-

Table 1

<table>
<thead>
<tr>
<th>Experiment</th>
<th>$T_s$ ($°C$)</th>
<th>$T_m$ ($°C$)</th>
<th>$\sigma_1$ (Pa)</th>
<th>$\sigma_2$ (Pa)</th>
<th>$\rho_1$ (g/cm$^3$)</th>
<th>$\rho_2$ (g/cm$^3$)</th>
<th>$\rho_3$ (g/cm$^3$)</th>
<th>$H_l$ (cm)</th>
<th>$H_{c1}$ (cm)</th>
<th>$H_{c2}$ (cm)</th>
<th>$h_f$ (cm)</th>
<th>$V$ (m/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>40</td>
<td>42</td>
<td>13</td>
<td>38</td>
<td>0.5</td>
<td>1.03</td>
<td>0.86</td>
<td>0.86</td>
<td>1</td>
<td>12</td>
<td>8</td>
<td>2</td>
</tr>
<tr>
<td>2</td>
<td>40</td>
<td>42</td>
<td>13</td>
<td>38</td>
<td>0.5</td>
<td>1.03</td>
<td>0.86</td>
<td>0.86</td>
<td>1</td>
<td>12</td>
<td>8</td>
<td>2</td>
</tr>
<tr>
<td>3</td>
<td>39</td>
<td>42</td>
<td>18</td>
<td>53</td>
<td>0.7</td>
<td>1.03</td>
<td>0.86</td>
<td>0.86</td>
<td>1</td>
<td>12</td>
<td>8</td>
<td>2</td>
</tr>
</tbody>
</table>

$T_s$ and $T_m$ are the temperatures at the surface and the base of the lithospheric model before deformation. $\sigma_1$, $\sigma_2$, $\sigma_3$ are the average values of yield limit for the mantle and upper and lower crustal layers of the lithosphere under normal load, respectively. $\rho_1$, $\rho_2$, $\rho_3$ are the densities of the mantle lithospheric layer, asthenosphere, upper and lower crustal layers, respectively; $V$ is the rate of the plate convergence; $H_l$, $H_{c1}$, $H_{c2}$ are the thickness of the mantle lithospheric layer, upper and lower crustal layers ($H_{c2}$ decreases to $h_f$ toward the subducting plate front, see Fig. 1).
tinent collision. In the central Himalayas the arc was of Andean-type [25], while in the NW Himalayas, the Ladakh and Kohistan arcs were intra-oceanic [26,27]. We do not aim to discuss in this paper the arc continent collision along the Himalayas and simply show in Fig. 9a a possible consequence of this process: the core (mainly mantle) part of the fore arc subducted into the mantle. During subduction of the Indian margin under this block the margin upper layers were partly scraped off and accreted to the overriding lithosphere at different depths within the interplate suture zone. A large part of the continental Indian crust, however, subducted to 200–250 km depth, following the Tethys oceanic lithosphere (Fig. 9a).

The crust then failed at several tens to about 100 km depth and rapidly moved upward between the two plates (Fig. 9b). The upward-moving crustal slice pushed up previously underplated crustal/sedimentary material squeezed in the interplate zone and now draping gneiss domes pervasive in the western Himalaya [28] and possibly represented among culminations of the South Tibetan North Himalayan belt [29]. Rapid uplift then stopped and was followed by slab break-off (Fig. 9b).

Subduction then switched to a highly compressional mode with tectonic underplating of the whole Indian lithosphere beneath Asia that underwent a first isostatic uplift (Fig. 9c,d), apparently in the Eocene, when oceanic sedimentation ceased on the Indian lithosphere (e.g. [30]) and residual basins persisted along the suture. The underthrust Indian crust was heated and thus softening. The physical state, the rheology and the mineralogical composition of this crust are not clear: Part of the crust, more likely the lower crust, was eclogitised and became denser than the mantle. The upper crust has likely been molten. After a few to several hundreds kilometers of tectonic underplating, the mantle layer (probably together with the lower crust) started to delaminate from the crust and to peel-back to the subduction zone (Fig. 9d). This process resulted in a new phase of isostatic uplift of the overriding plate (Tibet) and was terminated by a second break-off (Fig. 9e).

A second crust failure occurred in front of the subduction zone/mountain belt (Fig. 9e, highly compressional subduction regime). The major thrust fault formed at this time was the main central thrust (MCT). The low-viscosity Indian crust below Asia, in direct contact with the hot asthenosphere, became less viscous and started to intrude the hot dense Asian lithospheric mantle (Fig. 9e).

This southward propagating process (Fig. 9f) is consistent with mantle-derived heat input inferred to promote late Oligocene-early Miocene crustal anatexis in the arc region (e.g. [29,31]) and along the Himalayas [32,33–35]. The new subduction front was the MCT. Underplating along this fault further increased the crustal thickness below the Himalayas and resulted in the formation of high, intensively eroded relief. Erosion triggered both the exhumation of the Crystalline Himalayas and the formation of the South Tibetan normal fault system [36,37]. Subduction of the Indian lithosphere was again horizontal and resulted in further thickening of the crust under Tibet by intensively deforming its very weak (molten?) lower part and by adding new portions of the Indian crust.
Fig. 9. Evolutionary model for the Himalaya–Tibet system.

(a) Subduction of the Indian continental crust to \( \sim 250 \) km depth, scraping of the sedimentary and upper crustal material from the subducting plate, formation of the huge accretionary prism. Failure of the subducted crust at depth of ca. 100 km.;

(b) Rapid uplift of the subducted crustal slice to a few to several tens of kilometers depth. Break-off of the Indian mantle with the attached previously subducted oceanic lithosphere;

(c) Heating, weakening and uplift of the remaining in the mantle crustal segment of the Indian margin;

(d) Underplating of the Indian continental lithosphere under Asia and initiation of delamination of the Indian lithospheric mantle;

(e) Failure of the Indian crust in front of the orogen and initiation of the MCT;

(f) Replacement of the Asian mantle by the underplated Indian crust started at stage (e);

(g) Formation of the South Tibet detachment (STD) and exhumation of the metamorphics in the Crystalline Himalayas.

(h) Present stage with main boundary thrust (MBT).

1: Indian upper (a) and lower (b) crust; 2: Asian lithosphere: (a) continental crust, (b) lithospheric mantle; 3: scraped off and accreted Indian margin; 4: erosion; 5: thrust (a) and normal (b) faults.
The Asian lithospheric mantle (if any) was completely replaced by the Indian crust (Fig. 9g) and sank into the asthenosphere. This caused additional uplift of Tibet and Miocene crustal-derived leucogranites in the Himalayan Belt [33,34]. Both thickening of the Himalayan crust from below and erosion from above caused progressive exhumation of high-pressure/low-temperature rocks uplifted from ca. 100 km to shallow levels during earlier stages (Fig. 9b). Exhumation first occurred in areas where uplift and/or erosion rates were fastest, apparently at the western termination of the Himalayas, where high-pressure (HP) rocks are preserved [38-40]. In the tectonically similar North Himalayan domes the material scraped from the Indian margin and accreted at shallow depths has been pushed up, but the HP rocks are not yet exposed, or are not preserved.

The present stage (Fig. 9h) corresponds to the INDEPTH seismic profile [5]. In this figure we indicate the MBT as the major active thrust since ca. 10 Ma [41] merging on the crustal scale with the MCT. The formation of the MBT does not correspond to new failure of the whole Indian crust; it is a splay fault resulting from scraping off and accretion of the crustal/sedimentary material under and in front of the MCT.

6. Discussion

The amount of convergence in the model since 55 Ma is 1000 to 1500 km, which corresponds to the available estimates (see [42] for review). Subduction of so much continental lithosphere is multiphasic and includes the following major events: two break-offs (Fig. 9b,e); one delamination (Fig. 9d,e) and probably the onset of a second delamination (Fig. 9h); two rapid uplifts of the subducted crustal slices (Fig. 9b,g).

We do not anticipate that natural subduction occurred exactly in the same way, synchronously passing through the same stages over the >2500 km long Himalayan belt. The scenario proposed in Fig. 9 applies mostly to the Central Himalayas and is in agreement with recent tomography evidence for two pieces of high velocity material under India (Fig. 10a) interpreted as subducted and detached lithosphere [4]. The deepest portion could correspond to the Tethys oceanic lithosphere detached some 45 Ma ago (Fig. 9b) and the shallower one to the Indian lithospheric mantle detached ca. 25 Ma ago (Fig. 9e). A roll-over geometry of the subducted lithosphere with the deeper portion overturned and dipping southwards (Fig. 10) can be explained by the fact...
that Asia was apparently shortened NS by about 1000 to 1500 km since continental subduction started [43,44]. The subduction in front of the Himalayas has been respectively shifted over this distance to the north with respect to the mantle which is supposed to be fixed.

The tomography section across the western Himalayas displays only one piece of detached lithosphere (Fig. 10b), which could be identified as the Tethys slab. The vertical Indian lithospheric mantle is not detached and is bent to the south at 670 km depth transition zone. It is unclear whether this layer has first underplated Asia and then delaminated and peeled-back to its present position before the forthcoming break-off, or the separation between the mantle and crustal layers occurred as in exp. 2 (Fig. 6).

According to the model crustal thickening and uplift of Tibet was heterogeneous and first propagated to the north during tectonic underplating of the Indian crust under Asia (Powell and Conaghan [45] and Fig. 9c,d). Underplating should have started after a rapid rise of the HP rocks and break-off (Fig. 9b) some 45 Ma ago. Delamination and peel-back of the Indian lithospheric mantle (Fig. 9d,e) caused southward propagating uplift amplified by the replacement of the Asian lithospheric mantle by the Indian crust (Fig. 9f). This process very schematically presented in Fig. 9f is hypothetical. Its possibility and style (mechanism) should be tested numerically. Rise of the overriding plate continued due to further tectonic underplating of the Indian crust (Fig. 9h). Thus, Tibet uplift occurred through the whole Himalayan collision, but the most important pulse was Early Miocene (see Fig. 9d–f) as indicated by geological records (see [22] for review). According to the model this period corresponds to the formation of the MCT (Fig. 9e), which is not reliably dated but it is generally agreed that thrusting along this fault did not begin before the late Oligocene to early Miocene (e.g. [46]). Normal faulting along the STD (Fig. 9g) started approximately at the same time or somewhat later [47,48].

The proposed model is two-dimensional and has an undeformable overriding plate. The model is thus not designed to explain the Asian tectonics. Nevertheless, it provides ideas about the evolution of the effective strength and stress regime in the Asian lithosphere. The first conclusion that stems from modelling is that the strength of this lithosphere decreased from stage Fig. 9e until now due to thickening of its weak crustal layer and removal of the mantle layer (Fig. 9f). The NS compression within the Asian plate evolved as follows: initial continental subduction (Fig. 9a,b) was characterised by low-compressional regime and therefore compression of Asia was small. After break-off (Fig. 9c), 40–45 Ma ago the regime switched to a highly compressional mode and, therefore, the Asian plate was subjected to forceful compression, but was probably still too strong to fail. Compression then gradually increased during tectonic underplating of the Indian lithosphere and at 30–35 Ma became sufficient to cause major failure within the Asian lithosphere, leading to the formation of major strike-slip faults (e.g. the Aliao Shan Fault) [49,50]. Starting from stage Fig. 9e (20–25 Ma), the Asian lithosphere became considerably weaker while the subduction regime remained highly compressional until now. The intense deformation and failure of different modes (thrust and strike-slip) of this lithosphere is thus very likely. Indeed, along with the mentioned events (uplift of Tibet, formation of MCT and STD etc.) the 20–25 Ma period is characterised by the initiation of the major Asian faults such as the Altn-Tagh and Kun-Lun [9]. This is also the time of lithospheric-scale failure and initiation of intracontinental subduction in the Pamir, Qilian Shan and Tian-Shan (e.g. [51]).

7. Concluding remarks

The geopoem is consistent to a first approximation with geological and geophysical data. Although this model looks complex (Fig. 9), it is a very simplified two-dimensional representation of what may have happened in the Himalayas. Thermo-mechanical modelling shows to which extent the available data are limited and they are far from being sufficient to reconstruct the geodynamic evolution of a mountain belt over
a long time. This evolution occurs on the lithospheric scale and involves complex interaction between all lithospheric layers, which cannot be addressed without geodynamic modelling. Modeling, however, remains limited by the complexity and ambiguity of initial and boundary conditions. A further step in the modelling of continental subduction in the Himalayas consists in incorporating three dimensionality with deformable Asia.

Acknowledgements

We thank R. Van der Voo for providing unpublished tomography data, and C. Burchfiel and M. Harrison for the constructive review. This work was supported by 2HIM Himalaya (CNRS/INSU) programme. This is contribution number 274 of Geosciences Azur.[RV]

References