1 **Continental collision with a sandwiched accreted terrane: insights into Himalayan-**
2 **Tibetan lithospheric mantle tectonics?**

3 Sean Kelly¹, Jared P. Butler¹ and Christopher Beaumont¹

4 ¹Oceanography Department, Dalhousie University, Halifax, NS, Canada

5 Corresponding author:

6 Jared P. Butler
7 jaredbutler@dal.ca
8 Department of Oceanography
9 Dalhousie University
10 1355 Oxford Street
11 PO BOX 15000
12 Halifax, Nova Scotia
13 Canada, B3H 4R2

14
15 **Keywords**

16
17 Tectonics; geodynamic modeling; accreted terrane(s); continent–continent collision;
18 delamination; Himalaya-Tibet

19
20
21 **Abstract**

22
23 Many collisional orogens contain exotic terranes that were accreted to either the
24 subducting or overriding plate prior to terminal continent-continent collision. The ways in
25 which the physical properties of these terranes influence collision remain poorly
26 understood. We use 2D thermomechanical finite element models to examine the effects of
27 prior ‘soft’ terrane accretion to a continental upper plate (retro-lithosphere) on the ensuing
28 continent-continent collision. The experiments explore how the style of collision changes in
29
30
31
32
33
34

response to variations in the density and viscosity of the accreted terrane lithospheric mantle, as well as the density of the pro-lithospheric mantle, which determines its propensity to subduct or compress the accreted terrane and retro-lithosphere. The models evolve self-consistently through several emergent phases: breakoff of subducted oceanic lithosphere; pro-continent subduction; shortening of the retro-lithosphere accreted terrane, sometimes accompanied by lithospheric delamination; and, terminal underthrusting of pro-lithospheric mantle beneath the accreted terrane crust or mantle. The modeled variations in the properties of the accreted terrane lithospheric mantle can be interpreted to reflect metasomatism during earlier oceanic subduction beneath the terrane. Strongly metasomatized (i.e., dense and weak) mantle is easily removed by delamination or entrainment by the subducting pro-lithosphere, and facilitates later flat-slab underthrusting. The models are a prototype representation of the Himalayan-Tibetan orogeny in which there is only one accreted terrane, representing the Lhasa terrane, but they nonetheless exhibit processes like those inferred for the more complex Himalayan-Tibetan system. Present-day underthrusting of the Tibetan Plateau crust by Indian mantle lithosphere requires that the Lhasa terrane lithospheric mantle has been removed. Some of the model results support previous conceptual interpretations that Tibetan lithospheric mantle was removed by convective coupling to the pro-lithosphere. They can also be interpreted to suggest that delamination beneath Tibet was facilitated by densification and weakening of the plateau lithosphere, perhaps owing to long-lived pre- to syn-collisional subduction-related metasomatism beneath the Asian margin.

1. Introduction: Orogenesis involving accreted terranes

Continent-continent collisional orogenesis is commonly preceded by the accretion of one or more terranes to one or both of the continents as part of the Wilson Cycle (Wilson, 1966; Dewey et al., 1969). Accretion ranges from that seen in accretionary orogens (Cawood et al., 2009), comprising mainly stacked fragments derived from the subducting ocean, to small and large accreted terranes, either ‘rootless’ or attached to their lithospheric mantle.

Accreted terranes may be small (Williams and Hatcher, 1982; Van Staal et al., 1998) or large, e.g., those comprising the Tibetan Plateau (Dewey et al., 1988; Yin and Harrison, 2000). Accretion to the continent may be ‘soft’ or ‘hard’, the former involving ‘docking’ but limited tectonic shortening, overthrusting and metamorphism, and the latter involving structural emplacement of the allochthonous terrane, major tectonic thickening of crust and possibly mantle lithosphere, and metamorphism, leading to orogenesis within the accreted terrane itself. The terranes themselves include oceanic arcs, large igneous provinces, distal rifted margins, and ribbon and micro-continents (Roberts, 2003; Cawood et al., 2009; Tetreault and Buiter, 2014; Moresi et al., 2014). Moreover, they may be accreted while hot or cold, giving them a range of strengths depending on their prior history.

Here we present numerical models motivated by the question, ‘What effect does prior terrane accretion have on subsequent continent-continent collision?’ The case investigated is simple (Fig. 1)–that of a medium-sized continental lithospheric terrane which has undergone soft accretion to the upper plate (‘retro’-side) of the system. Using 2D thermomechanical finite element model experiments we explore how the style of orogenesis varies with the properties of the pro-continental and accreted terrane lithospheric mantles. In particular, models are designed to assess:

1. how oceanic slab break-off influences the subsequent subduction of the pro-continent mantle;
2. under what circumstances subduction of pro-continent mantle under the accreted terrane leads to advancing subduction and shortening of the accreted terrane;
3. how tectonic shortening of the accreted terrane is accommodated by its lithospheric mantle; and,
4. the influences of crustal thickening and lithospheric mantle deformation/delamination on the topographic and thermal evolution of the orogen.

Even with a simple model design (Fig. 1), many variables could be considered. We purposely confine our investigation to factors we consider first-order in regard the behavior of the lithospheric mantle, namely the density of the pro-continent mantle and the density and viscous strength of the accreted terrane lithospheric mantle. We justify this

choice by appealing to the Archimedes number for the mantle lithosphere which determines its propensity to deform and/or subduct (Appendix A2).

While the model design is intended to be a generic prototype for similar natural systems, the results reveal emergent behaviors similar to those previously diagnosed for the more complex Himalayan-Tibetan (H-T) orogen. We therefore conclude with a discussion of the applicability of our results to the H-T system, focusing on the controversial problem of how India–Asia convergence was accommodated at the lithospheric scale (e.g., Argand, 1924; Dewey et al., 1988; Le Pichon et al., 1992; Molnar et al., 1993; Willett and Beaumont, 1994; Owens and Zandt, 1997; Shen et al., 2001; Tapponnier et al., 2001; Johnson et al., 2002; Royden et al., 2008; van Hinsbergen et al., 2011a).

2. Modeling methods

2.1 Model design

We use the 2D finite element software SOPALE-nested to solve thermomechanical creeping flows at the mantle scale (Fullsack, 1995; Beaumont et al., 2009; Butler et al., 2014, details Appendix A1). The starting geometry of the models (Fig. 1a; Table 1) represents a simplified continent-continent collisional system comprising (left to right) a pro-continent (P), an ocean, and a composite retro-lithosphere, overlying upper mantle to depth of 660 km and lower mantle to 1200 km. The retro-lithosphere contains a 900-km-wide accreted continental terrane (AT), and an interior, older stronger retro-continent (R). There is no weak zone between them. Continents are 120 km thick (4km sediments, 20 km upper crust, 12 km lower crust and 84 km lithospheric mantle). Oceanic lithosphere is 100 km-thick (2 km sediments, 8 km crust and 90 km lithospheric mantle) corresponding to a 100 Myr-old oceanic plate. A small, pre-existing accretionary wedge is situated between the oceanic plate and the retro-lithosphere accreted terrane below which a thin weak dipping layer of oceanic crust embedded in the oceanic lithospheric mantle acts to initiate subduction.

Model materials deform by either frictional-plastic (brittle) or power-law viscous (ductile) flows that include the effects of strain-softening/strain-weakening mechanisms, respectively. At high stress, a saturation stress approximation to the Peierls (high stress) flow law modifies the power-law flow. Frictional-plastic deformation depends on the

dynamical pressure and effective angle of internal friction (ϕ_{eff} ; hereafter simply ϕ). For all model materials except the sediments, ϕ varies from $15-2^\circ$, (15° is the effective angle of internal friction, and 2° is fully strain-softened material). Sediments are initially weaker, with $\phi = 8-2^\circ$. Strain-softening is achieved by linearly decreasing ϕ over the range of strain $\varepsilon = 0.5-1.5$ (Appendix A1).

Viscous deformation is calculated based on a small set of reliable, laboratory-derived flow laws (Table 1). Where desired, the viscosity is scaled by a factor f in order to represent materials stronger or weaker than the base flow law (Appendix A1; see justification in Butler et al. 2014). During viscous strain-weakening, the effective viscosity is linearly scaled between 1 and $1/W_s$ over the range of strain $\varepsilon = 5-10$ (Table 1, Appendix A1). The difference in the range of strain for softening and weakening is chosen to reflect onset of strain localization and faulting in brittle rocks at low levels of deformation versus the requirement that larger strains are needed to produce grain-size reduction, mylonitization and other weakening mechanisms during viscous flow. During strain-weakening continental upper crust follows a scaled wet quartzite flow law ($WQ \times f$; Gleason and Tullis, 1995), where f varies among crustal units. The pro-continent and accreted terrane crusts have WQ with $f = 1$, whereas the stronger retro-continent interior crust has WQ with $f = 10$. Lower continental and oceanic crusts have a scaled dry Maryland diabase flow law ($DMD \times 0.1$; Mackwell et al. 1998). Mantle has a wet olivine flow law (WOL; Karato and Wu, 1993) scaled to represent either sublithospheric mantle or dehydrated lithospheric mantle (see below). Owing to considerable uncertainty in the rheology of the lower mantle (Billen, 2010), we assume that the lower mantle has a constant viscosity of 1×10^{21} Pa s.

Upper continental crust melt weakens (Beaumont et al., 2001; 2004), such that the viscosity decreases linearly from the calculated power-law value to a fixed linear value of 10^{19} Pa s over the temperature range $700-750^\circ\text{C}$. This represents the effect of minor melting, melt fraction between 4 and 7% (Rosenberg and Handy, 2005).

The model domain has a stress-free upper surface, and free-slip sides and base. Convergence in the models is driven by pro-lithosphere boundary velocity, V_p , the retro lithosphere boundary is stationary (Fig. 1). $V_p = 15 \text{ cm a}^{-1}$ during oceanic subduction, consistent with paleomagnetic constraints on the rate of pre-collisional India-Asia

convergence (van Hinsbergen et al., 2011b), and decreases to 5 cm a⁻¹ during collision. The tectonic boundary force on the lithosphere required to achieve these rates ranges from about -2500 GN m⁻¹ during the oceanic subduction phase, indicating a net tensile stress resulting from slab pull, to between 1800 to 2400 GN m⁻¹ compression during the collision phase, depending on the properties of the pro-lithosphere and accreted terrane. In all models, the compressive force required to drive orogenesis remains lower than estimates of ridge push (typically 2500 GN m⁻¹). The model results for the post-collisional phase are insensitive to the value of V_p during oceanic subduction. Collision is defined as time of first contact between the pro-continent and accretionary wedge.

Thermal boundary conditions (Table 1) are basal heat flux 21 mW m⁻²; 0°C surface, and insulated side boundaries. Radioactive heat production, A_R , is respectively 2.0 and 0.4 μ W m⁻³ for upper and lower crust, resulting in an initial steady-state Moho temperature of 670°C, and 62 mW m⁻² surface heat flow for the continents, including the accreted terrane. Crustal materials undergo reversible pressure- and temperature-dependent density changes at the eclogite and coesite-eclogite phase transitions (Warren et al., 2008) (Appendix A1, Table 1). Density reversal implies retrogression, not necessarily to the protolith, but to a material with the original density, e.g., from eclogite to amphibolite (Warren et al., 2008). Mantle materials undergo an approximately 8% increase in density at the 660 km upper/lower mantle boundary, which is also treated as a reversible phase change boundary with a Clapeyron slope of -2 MPa K⁻¹.

The dynamically consistent model topography (Appendix A1) is a product of the model calculation without any imposed constraints except for submarine water loading. For simplicity there is no surface erosion.

2.2 Experimental Parameters

To first order, lithospheric mantle dynamics can be interpreted in terms of the non-dimensional Archimedes number, Ar , the ratio of gravitational forces to viscous forces (Appendix A2);

$$Ar^{1/2} = \frac{\rho_{\text{eff}} g L^3}{\eta V L}$$

For the lithospheric mantle, the effective density is the density difference between the sublithospheric and lithospheric mantles, $\rho_{\text{eff}} = (\rho_m - \rho_l)$, g is the gravitational

acceleration, L , η and V are characteristic length, viscosity and velocity scales, the last being the convergence boundary velocity. This form of the Archimedes number is appropriate when there is a defined characteristic velocity. In regard to lithospheric mantle processes, e.g. subduction or delamination, the main governing model properties for a fixed V are therefore the effective density and viscous strength of either lithospheric or sublithospheric mantle. The effective density has two components, the compositional density difference at Standard Temperature and Pressure, and the differential effect of thermal expansion owing to a colder mantle lithosphere than sublithospheric mantle. The thermal expansion effect is the same in all models, therefore, we only refer to the compositional component in the model description.

We chose a prototype model and a set of 36 model experiments based on variations of the prototype. These variations are designed to explore how the style of continent–continent collision is influenced by three quantities: density and viscosity of the accreted terrane lithospheric mantle, and density of the pro-continent lithospheric mantle. Variations in the density of the lithospheric mantle are expressed in terms of their effective density with respect to the sublithospheric mantle ($\Delta\rho = \rho_{\text{eff}} = \rho_m - \rho_l$), and can be interpreted in terms of the effects of chemical depletion (varying Mg#) on the density of peridotite (Lee, 2003; Griffin et al., 2009). We therefore refer to the effective density as the depletion density, where a smaller lithospheric mantle density, i.e., larger depletion density, reflects greater depletion (Table 1). In the prototype model, the accreted terrane lithospheric and pro-lithospheric mantles have respective depletion densities of $\Delta\rho = 50 \text{ kg m}^{-3}$ and 25 kg m^{-3} . In the sensitivity experiments, we tested a $\Delta\rho$ of 0, 10, 20 and 40 kg m^{-3} for the accreted terrane lithospheric mantle, and 30, 40, and 50 kg m^{-3} for the pro-continental lithospheric mantle (Fig. 1b).

The role of the accreted terrane lithospheric mantle viscosity was tested by varying the viscosity scaling factor f applied to the wet olivine flow law, from 5 to 3.5, and $2\times\text{WOL}$, where 5 represents a moderately dry mantle and 2 represents wet mantle, identical to sublithospheric mantle. When combined, the density and viscosity can be likened geologically to fertility of the accreted terrane lithospheric mantle. The undepleted, low viscosity ($f=2$) state represents fertile mantle that has not undergone melt

extraction/dehydration, or alternatively has been re-fertilized/hydrated by mantle metasomatism.

3 Prototype model result

3.1 Model AT-PR: pro-lithosphere subduction with negligible retro-lithosphere deformation

We describe all models in terms of the depletion density ($\Delta\rho$) and f -scaling (f), of their pro-lithosphere (P) and accreted terrane (AT). The prototype model (Model AT-PR; Table 1) is therefore P($\Delta\rho = 25; f = 10$)-AT($\Delta\rho = 50; f = 5$) meaning it has a relatively high-density pro-lithospheric mantle, $\Delta\rho = 25 \text{ kg m}^{-3}$, chosen to illustrate a case which favors subduction of the pro-lithosphere mantle with limited subduction zone (S-point) advance. The accreted terrane mantle lithosphere is highly depleted ($\Delta\rho = 50 \text{ kg m}^{-3}$) and strong ($f = 5$), representing a depleted/not metasomatized end-member that resists removal. When describing other models, we omit $\Delta\rho$ and f . Model times are reported in millions of years, post-collision (Myr-pc).

At 0-Myr pc (not shown), the oceanic lithosphere has subducted and the leading edge of the pro-continent margin contacts the accretionary wedge. The pro-continent now converges at 5 cm a^{-1} . By 2.5 Myr-pc (Fig. 2a) the leading edge of the pro-continent margin has subducted beneath the accreted terrane ultra-high-pressure metamorphic conditions. The negative buoyancy of the oceanic slab overcomes its strength and the slab breaks off, sinking deep into the upper mantle. By 14 Myr-pc (Fig. 2b) the system has undergone 750 km of convergence since collision, accommodated by accretion and thickening of a ~ 400 km-wide region of pro-continental crust, to a maximum thickness of ~ 70 km adjacent the suture. UHP pro-continent material exhumed to the base of the pro-side orogenic middle crust, driven by its buoyancy (Fig. 2b, blue adjacent to the suture). This behavior is typical of many of the models described here, and is similar to the exhumation processes seen in previous models of UHP rock exhumation during early continent-continent collision (Warren et al., 2008; Beaumont et al., 2009; Butler et al., 2013). The advance of the suture (the contact between the pro-continent and accreted terrane crusts) and S-point (the point at which the subducting lithosphere descends from the overlying crust/lithosphere) are not significant (Fig. 2b, S and SP); therefore, the accreted terrane remains relatively undeformed. Convergence continues to be accommodated in this fashion and by 30 Myr-pc

(Fig. 2c) the orogen has developed a large, ~900-km-wide region of detached and thickened pro-crust backed by relatively undeformed accreted terrane lithosphere. Within the thickened pro-crust, thermal relaxation and radioactive heating have raised the temperature of the Moho to 600–700°C in the vicinity of the suture. Beneath the orogen, the relatively buoyant pro-continent sublithospheric mantle has adopted a shallower subduction trajectory. At 45 Myr-pc (Fig. 2d) the orogen is >1300 km wide, and comprises only accreted pro-continent crust, the base of which has undergone further heating to ~800°C in response to radioactive heating and thermal relaxation. Cumulative shortening of the retro-lithosphere is ~80 km as indicated by advance of the suture, but there is little shortening/thickening of the accreted terrane crust and its underlying lithospheric mantle remains largely undeformed.

In summary, model AT-PR represents a case where the pro-lithosphere is relatively dense (low level of depletion) and tends to subduct once the crust has detached. The weak tendency for subduction zone advance is resisted by the viscously strong mantle lithosphere of the accreted terrane and its highly depleted, buoyant nature.

4. Model results from the Sensitivity analysis

4.1 Summary of model results

The results of all 36 sensitivity model experiments (Fig. 3) demonstrate a contrast with the prototype model AT-PR (Fig. 2) owing to the difference in pro-lithosphere mantle depletion density, 30–50 kg m⁻³ versus 25 kg m⁻³. Increasing depletion density makes the pro-lithosphere more buoyant and increases its propensity to underthrust the orogen as a flat slab. This difference is responsible for the advancing subduction nature of the sensitivity models (Fig. 3). It is also responsible for the entire shortening of the retro-side of the orogen in contrast to model AT-PR where there was no deformation. This contrasting behavior indicates that advancing subduction strongly correlates with, and may require relatively buoyant pro-lithosphere mantle. Advancing subduction, the opposite of subduction zone (or trench) retreat, is a requirement in orogens like the Himalayan-Tibetan orogen where India has both indented Asia and subducted beneath it. Advancing subduction also leads to trailing of the subducted pro-lithosphere mantle slab behind the advancing S-point, seen as curling under of the slab in all of the sensitivity models (Fig.3).

Initially, all of these continental mantle slabs are negatively buoyant and sink despite having a depletion density of 30–50 kg m⁻³. However, as the slabs heat up they become buoyant (Appendix Fig. S2, density plot) and rise beneath the overlying pro-lithosphere, either adopting a folded-back geometry (Fig. 4) for the most depleted cases, or curled geometry (Fig. 5), where there is less depletion (Fig. 3). The subducted slabs have an initial average temperature of 975 °C and for the properties used in these models the 50 kg m⁻³ slab is almost neutrally buoyant, whereas those with 30 and 40 kg m⁻³ depletions only require heating by an average of approximately 150 and 70°C, respectively, to become neutrally buoyant. We suggest that either the folded or curled case could be a possible explanation where tomographic images show regions of apparently thick continental lithospheric ‘lumps’ beneath large collisional orogens that were subject to advancing subduction.

The styles of the sensitivity model experiments (Fig.3) can be subdivided into four main regimes, with emphasis on the style of lithospheric mantle deformation. We use the terms northward and southward to refer to ‘directed to the right’ or ‘left’, respectively in the Figure 1 configuration.

1. Type 1 models (coded green) produce efficient shortening of the accreted terrane crust accompanied by rapid delamination and sinking of the accreted terrane lithospheric mantle, usually involving its northward subduction.

2. Type 2 models (coded red) similarly produce efficient shortening of the accreted terrane crust, but in this case accompanied by southward delamination of the accreted terrane lithospheric mantle and removal by convective entrainment by the subducting pro-continent lithosphere.

3. Type 3 models (coded purple) produce shortening of the accreted terrane crust accompanied by lateral shunting/translating the buoyant accreted terrane mantle lithosphere beneath the retro-lithosphere.

4. Type 4 models (coded blue) produce relatively minor shortening of the accreted terrane crust and the buoyant accreted terrane mantle lithosphere is squeezed and sheared, ‘smeared’, but there is no wholesale delamination or shunting.

Below we describe the results of representative Type 1–4 models which then serve as a template for other models of the same type shown in Figure 3. Comments on the distribution of these types within the experimental set are given in Figure 3 (caption).

4.2 Model Type 1: Delamination of AT mantle by retroward (northward) subduction

Model AT-1, P(50:10)-AT(0:2) (Fig. 4), is an opposite end-member to Model AT-PR in that the pro-continent mantle lithosphere is highly depleted and the accreted terrane lithospheric mantle has a depletion density of 0 kg m^{-3} and a viscosity scaling factor of $f = 2$, and thus represents the most fertile/metasomatized end-member of the P(50:10) experiment set (Figs. 1 and 3c).

As in Model AT-PR, at 2 Myr-pc (Fig. 4a) the leading edge of the pro-continent crust has subducted to UHP conditions beneath the retro-lithosphere, and the oceanic slab has broken off and begun sinking into the mantle. Subsequent convergence involves detachment and thickening of both pro- and accreted terrane crusts. By 7.5 Myr-pc, detachment of accreted terrane crust has formed a retro-wedge. With further convergence, shortening in the accreted terrane jumps to its northern boundary, resulting in the formation, by 13.5 Myr-pc (Fig. 4b), of two oppositely-verging crustal wedges at either end of the terrane. This crustal shortening accompanies deformation of the underlying lithospheric mantle, which has thickened and warped downward at its ends, becoming entrained with the subducting pro-lithosphere at its southern end. Lithospheric mantle at its northern end begins to sink into the sublithospheric mantle. Continued shortening forms a widening orogen with a high plateau, comprising both detached pro-continent and accreted terrane crusts (Fig. 4c). Beneath the orogen, the gravitational instability at the northern end of the accreted terrane first drips and then develops into wholesale delamination and northward subduction of the accreted terrane lithospheric mantle. By 20.5 Myr-pc (Fig. 4c) the latter peels away rapidly from north to south, sinking into the sublithospheric mantle. During delamination the base of the accreted terrane crust is exposed to upwelling sublithospheric mantle, which heats the Moho from $\sim 700^\circ\text{C}$ to $\sim 1000^\circ\text{C}$ in just a few million years. Upwelling of the hot sublithospheric mantle also leads to rapid uplift of the north end of the plateau, which increases in elevation by $\sim 1\text{--}1.5 \text{ km}$ during delamination, between 18 and 20.5 Myr-pc (Fig. 4c, inset). At the time of delamination accreted terrane shortening has accommodated $\sim 600 \text{ km}$ of suture zone (S)

and S-point (SP) advance with attendant hinging and curling under of the trailing subducted pro-continent mantle (Fig. 4c); trailing being a consequence of advancing subduction. Subsequent convergence sees this buoyant pro-continent lithospheric mantle underthrust the plateau. By 26.5 Myr-pc (Fig. 4d) the latter has underthrust to the northern end of the accreted terrane, thereby closing the ‘mantle window’ opened by delamination. The final orogen comprises an 800-km-wide plateau consisting of both thickened pro-continent and accreted terrane crust, underlain by flat-slab subducting pro-lithospheric mantle. In contrast to Model AT-PR, the flat-slab portion of the pro-continent lithosphere now directly underlies the full width of the plateau crust. The combined effects of thermal relaxation, radioactive heating, and exposure of the orogenic crust to hot sublithospheric mantle have heated the Moho to $\sim 800^{\circ}\text{C}$, with 900°C at the base of the thickened upper crust at the northern end of the accreted terrane.

4.3 Model Type 2: Delamination of AT mantle by pro-continent entrainment

In Model AT-2, P(40:10)-AT(0:2) (Fig. 5), the pro-continent lithospheric mantle has a depletion density of 40 kg m^{-3} , marginally less than the 50 kg m^{-3} in Model AT-1. Consequently, its early evolution is similar. By 7.5 Myr-pc, pro-continent crust has detached, forming an orogenic pro-wedge. Some of this crust has again been metamorphosed at UHP conditions and exhumed to the base of the wedge. Compared with Model AT-1, the accreted terrane crust is less deformed and a greater amount of convergence has been absorbed by subduction of pro-continental lithospheric mantle, consistent with the latter having a higher density in Model AT-2. Continued convergence leads to the development of a broad plateau comprising detached pro-continent and thickened accreted terrane crust. Later, the accreted terrane lithospheric mantle becomes entrained by the advancing subducting pro-continent lithosphere, resulting in ‘double subduction’ (Fig. 5a; 24 Myr-pc; Supplementary Material Figs S1–S3). Pro-lithosphere subduction now outpaces the rate of subduction zone (S-point) advance (Fig. 5a, b, SP) and beginning ~ 28 Myr-pc, the remaining accreted terrane lithospheric mantle decouples from the crust and is rapidly swept southward by convective entrainment, leaving a mantle window in its wake. Sublithospheric mantle upwells in response and by 32 Myr-pc (Fig. 5c), the entire accreted terrane lithospheric mantle has been removed, and close to a half of the orogenic plateau is now underlain by hot sublithospheric mantle. As in Model AT-1, this

leads to dramatic heating of the now exposed Moho, from $\sim 700^{\circ}\text{C}$ to $\sim 1100^{\circ}\text{C}$. Rapid uplift ensues, with the plateau gaining about 0.5–1 km of elevation within a few million years (Fig. 5c, inset). During continued convergence the pro-continent lithosphere mantle begins to underthrust the accreted terrane crust, but does so less rapidly than in Model AT-1, owing to its reduced buoyancy. Nonetheless, by 50 Myr-pc (Fig. 5d) the accreted terrane lithospheric mantle is largely underlain by a flat slab of pro-continent lithospheric mantle, with only a minor mantle window remaining.

4.4 Model Type 3: Shunting of AT mantle lithosphere

Model AT-3 P(50:10)-AT(40:2) (Fig. 6) has a pro-continental lithospheric mantle depletion of 50 kg m^{-3} , and relatively depleted (40 kg m^{-3}) but weak ($f = 2$) accreted terrane lithospheric mantle. Its early evolution is similar to that of the above models, with oceanic slab breakoff, minor subduction of pro-continental crust to UHP conditions, and formation of a broad pro-wedge comprising accreted pro-continental crust (Fig. 6a; 14 Myr-pc). Advancing subduction and consequent shortening of the accreted terrane lithosphere begins before 14 Myr-pc, is well underway by 20 Myr-pc (Fig. 6b) and by 25 Myr-pc (Fig. 6c) the accreted terrane crust has doubled in thickness. The subducting pro-continent mantle has taken on a shallow, flat-slab trajectory, indenting and shortening the adjacent accreted terrane lithospheric mantle. At its northern end, the accreted terrane lithospheric mantle has started to underthrust the adjacent retro-lithospheric mantle. With continued convergence the accreted terrane lithospheric mantle is further squeezed and shunted beneath the retro-lithosphere, and by 31 Myr-pc (Fig. 6d) the shortened accreted terrane crust is half underlain by pro-continent lithospheric mantle. The accreted terrane lithospheric mantle does not sink owing to its 40 kg m^{-3} depletion density, which makes it buoyant. Instead, it spans the southern edge of the retro-continent lithosphere. Unlike Models AT-1 and AT-2 where the accreted terrane delaminates, there is no crustal exposure to hot sublithospheric mantle and thus no rapid, buoyant uplift of the plateau. Instead progressive thickening of the accreted terrane crust leads to laterally expanding plateau formation throughout the model evolution (Fig. 6d, inset).

4.5 Model Type 4: Smearing of the AT mantle lithosphere

Model AT-4 P(40:10)-AT(20:5) (Fig. 7) has a pro-continent lithosphere depletion of 40 kg m^{-3} , and depletion of 20 kg m^{-3} and $f = 5$ in the accreted terrane mantle, placing it among

the strong end-members with small depletion density. Its early evolution is similar to that of Model AT-2, with steep subduction of the pro-continental lithospheric mantle accompanied by formation of a widening crustal prowedge (Fig. 7a; 7.5 Myr-pc). Shortening continues in this mode for much of the model evolution, leading to the formation by 20 Myr-pc (Fig. 7b) of a ~600-km-wide detached and thickened pro-crust. As convergence continues the subducting pro-continent adopts a progressively shallower trajectory leading to underthrusting of the accreted terrane and curling of the trailing subducted mantle. By 30 Myr-pc (Fig. 7c) the strong accreted terrane has begun to shorten slowly, leading to a nascent orogenic wedge at its northern end, coupled with minor downwarping of the underlying mantle. By 43 Myr-pc (Fig. 7d) the accreted terrane has been shortened to ~630 km, leading to substantial, but non-uniform crustal thickening. The underlying strong mantle lithosphere has neither detached nor sunk, nor has it been shunted away, but has instead been smeared largely in place, between the thickened crust and the underthrusting pro-lithosphere mantle, with minor northward translation. Its strength retards underthrusting by the pro-lithosphere mantle, forcing it underneath such that the underthrusting more than doubles the thickness of mantle lithosphere beneath the accreted terrane (Fig. 7d). As in Type 3, there is no dramatic rapid uplift of the plateau, but rather a gradual laterally expanding uplift related to thickening of the pro-continent crust, and later lateral advection of the high topography as the suture advances, accompanied by uplift within the accreted terrane as it undergoes some shortening (Fig. 7d, inset).

5.0 Discussion 1

5.1 Previous numerical models of terrane accretion and advancing continent–continent collision

Most recent modeling studies have focused on terrane accretion (e.g., Moresi et al., 2014) rather than the ensuing continent-continent collision. Tetreault and Buiter (2014), for example, modelled accretion of small terranes, their FATs, analogous to continental fragments, island arcs, and large igneous provinces. Vogt and Gerya (2014) investigated subduction/accretion of oceanic plateaus. Butler et al. (2013, 2014) also explored accretion of a small micro-continental block in the context of ultra-high-pressure metamorphism. In these studies, terrane accretion involves off-scraping of the crust from the subducting

lithosphere, akin to formation of an accretionary wedge, rather than the soft collision style of accretion that we envisage as a precursor to the present models. Moreover, they do not consider how terrane accretion may pre-condition later continent-continent collision. A few previous studies explored the response of existing within-lithosphere lateral weak regions, akin to accreted regions, to compression (Ellis et al., 1998; Krystopowicz and Currie, 2013). However, these models did not include oceanic and continental subduction, making them more applicable to intra-continental orogenesis. Most recently, Sternai et al. (2016) modeled advancing continental subduction in 3D in the context of Himalayan-Tibetan orogenesis and conclude sublithospheric mantle flow may be a significant forcing factor. However, their approach does not address the consequences of prior terrane accretion *per se*.

5.2 Relationship to other mechanisms of lithospheric mantle removal

Numerical modeling studies have, however, investigated mechanisms potentially responsible for weakening and removal of all or part of the continental mantle lithosphere during shortening. These studies range from detailed calculations of viscous Rayleigh-Taylor instabilities (e.g., Houseman et al., 1981; Conrad and Molnar, 1997; Houseman and Molnar, 1997; Molnar et al., 1998), to models that explore delamination in response to lower crustal density increase through eclogitization during thickening (Sobolev and Babeyko 2005; Krystopowicz and Currie, 2013), or underplating of the crust by dense mafic melts (Elkins-Tanton, 2005). The role of lower crustal decoupling zones in producing lithospheric delamination/decoupling has also been explored (Morency and Doin, 2004; Le Pourhiet et al., 2006; Göğüş and Pysklywec, 2008).

Here we invoke different mechanisms, metasomatism (specifically melt metasomatism; O'Reilly and Griffin, 2013) that weakens and increases the density of mantle lithosphere through re-fertilization and rehydration, or the opposite effect of differing degrees of initial depletion. Unlike Wenker and Beaumont (in prep.), the metasomatism is not modelled explicitly. Instead, we consider the end product of different degrees of metasomatism acting on the accreted terrane, assuming that metasomatism predates continental collision. The models may therefore be interpreted in terms of metasomatism of accreted terrane mantle lithosphere that was initially strongly depleted, dehydrated and refractory (depletion density 50 kg m^{-3} and $f = 5$) and was subsequently

metasomatized to differing degrees, represented by the sensitivity analysis. On the other hand, the sensitivity analysis can also be interpreted to investigate cases where the initial depletion and dehydration varied and there was no subsequent metasomatism.

All of the sensitivity model results (Fig. 3) demonstrate a contrast with the prototype model AT-PR (Fig. 2) owing to the difference in pro-lithosphere mantle depletion density, 30–50 kg m⁻³ versus 25 kg m⁻³. Increasing depletion density makes the pro-lithosphere more buoyant and increases its propensity to underthrust the orogen as a flat slab. This difference is responsible for the advancing subduction nature of the sensitivity models (Fig. 3). It is also responsible for the entire shortening of the retro-side of the orogen in contrast to model AT-PR where there was no deformation. This contrasting behavior indicates that advancing subduction strongly correlates with, and may require relatively buoyant pro-lithosphere mantle. Advancing subduction, the opposite of subduction zone (or trench) retreat, is a requirement in orogens like the Himalayan-Tibetan orogen where India has both indented Asia and subducted beneath it.

Advancing subduction also leads to trailing of the subducted pro-lithosphere mantle slab behind the advancing S-point, seen as curling under of the slab in all of the sensitivity models (Fig.3). Initially, all of these continental mantle slabs are negatively buoyant and sink despite having a depletion density of 30–50 kg m⁻³. However, as the slabs heat up they become buoyant (Supplementary Material, Fig. S2, density plot) and rise beneath the overlying pro-lithosphere, either adopting a folded-back geometry (Fig. 4) for the most depleted cases, or curled geometry (Fig. 5), where there is less depletion (Fig. 3). We suggest that either case could be a possible explanation where tomographic images show regions of apparently thick continental lithospheric ‘lumps’ beneath large collisional orogens that were subject to advancing subduction.

The sensitivity models demonstrate four emergent types of deformation of the accreted terrane lithospheric mantle, namely: 1) pro- and retro-ward delamination, 2) entrainment by pro-continent, 3) shunting and, 4) smearing (Figs 4–7). These types reflect increasing difficulty of removal of the mantle from Types 1 and 2 (weak, $f=2$, and small depletion density lithosphere) to Types 3 and 4 (stronger $f=2, 3.5, 5$ or with a large depletion density, 40 kg m⁻³). The main point is that the mantle lithosphere may delaminate and sink (Figs 4, 5), be shunted aside (Fig. 6) or remain in place and enjoy being deformed (Fig. 7).

There is also a time dependence of the response with removal in Types 1 and 2 occurring earlier in the collision and Types 3 and 4 progressively later. There is no proof that these types represent realistic ways to remove mantle lithosphere, but they do offer possibilities that should be investigated as alternatives to the more classical dripping and rollback mechanisms. In all cases removal is initiated by the advancing subduction of the pro-lithospheric mantle which shortens the accreted terrane, thickens its crust, and destabilizes its mantle lithosphere.

The second main emergent property of the models is that removal of the accreted terrane lithospheric mantle facilitates underthrusting by the buoyant pro-lithospheric mantle; earlier advancing subduction changes to underthrusting (Figs 4–7). In Types 1-3 the accreted terrane mantle lithosphere is replaced by the underthrusting pro-lithospheric mantle, but for Type 4 underthrusting produces a composite double mantle lithosphere.

The models do not contain any special weak or dense layers in the lithosphere designed to promote delamination (e.g., Göğüş and Pysklywec, 2008) other than the parameter variations described above. In complementary model investigations, e.g., Krystopowicz and Currie (2013), transformation of lower crust into dense eclogite drives delamination. In our models lower continental crust transforms to eclogite but it is less dense than the sublithospheric mantle, owing to the assumed intermediate composition, (Table 1) and does not drive delamination (Figs 4–7). Making the eclogite denser would certainly promote delamination and sinking of the lower lithosphere, so that Type 1 and 2 behaviors would develop in some current Type 3 and 4 models.

6.0 Discussion 2: Application to the Himalayan-Tibetan orogeny

6.1 Preliminary comparison of Himalayan-Tibetan orogen with model results

We now consider whether the model results provide insight into the Himalayan-Tibetan (H-T) orogeny. The present models are simple. They neither reproduce the complexity nor scale of the H-T orogeny. Nevertheless, some of the type behaviors may provide insight, particularly in regard to the Lhasa terrane, the equivalent of our accreted terrane. Bearing in mind the uncertainties and widely ranging interpretations concerning the evolution and present structure of the H-T orogen (examples shown Fig. 8 and references in caption), we assess whether the models meet the more robust constraints at the lithospheric scale,

specifically in regard to advancing subduction and removal of lithospheric mantle, and whether their emergent behaviors support novel interpretations of existing data. This represents a short preliminary comparison with the H-T system intended to promote discussion, with more detailed analysis to follow in models with multiple accreted terranes.

We assume the interior of Asia (Siberia) has been nearly stationary since the H-T collision, which is consistent with paleomagnetic evidence (Dupont-Nivet et al., 2010; Lippert et al., 2014), and plate circuits (van Hinsbergen et al. 2011b). We therefore use the same reference frames for the H-T orogeny (stable Asia) and our models (stationary retro-boundary. We also assume collision of Tethyan Himalaya with the Lhasa terrane at approximately 55 Ma (Yin and Harrison, 2000; and refs. therein; Klemperer, 2006). The question of the significance of pre-collision deformation of the Lhasa terrane (e.g. Kapp et al., 2007; DeCelles et al., 2007) is not the main focus of the comparison but can be approximated by assuming a starting configuration like that at ca. 10–15 Myr-pc in the models, after some shortening and thickening of the accreted terrane, but prior to removal of its mantle lithosphere.

6.2 Northward translation of the ITS suture

Motion of the surface suture (the contact between the Indian and Asian crusts; ITS, Fig. 8) provides a crucial constraint on the evolution of the H-T system. Paleomagnetic data indicate the Lhasa terrane has migrated northward with respect to stable Asia by $\sim 1100 \pm 500$ km since collision (Dupont-Nivet et al., 2010; Huang et al., 2015) to its present $\sim 30^\circ\text{N}$ latitude in the central Tibetan Plateau (Yin and Harrison, 2000). Seismic tomography has identified the inferred remnant Neo-Tethyan oceanic slab, which presumably broke off shortly after collision, and now lies ~ 1000 km south of the present surface suture (Fig. 8b) (e.g., Van der Voo et al., 1999; Hafkenscheid et al., 2006; Replumaz et al., 2010c). Therefore, the suture has migrated north dramatically following collision owing to shortening of terranes between it and stable Asia, which is confirmed by India ‘indentation’ measured at the syntaxes.

Most of the present sensitivity models show a similar evolution to the H-T system. All exhibit break-off and sinking of the oceanic lithosphere shortly after collision, but it remains at the same horizontal position acting as a marker. The model surface suture (S) then advances retroward (right) of the detached slab. This advance, ~ 500 – 600 km in most

models (initial position marked by red arrow in Figs 4–7), is less than for the H-T system because only one accreted terrane is shortened. There are, however, two essential requirements for this behavior which relate to the H-T system. First, the pro-lithosphere must be sufficiently buoyant (chemically depleted), that it partly resists subduction and exerts a compressive force on the accreted terrane. This condition is met by all models in the parameter space explored here, except the prototype (Fig. 2). Second, the accreted terrane must be sufficiently weak that it deforms and allows retroward suture zone advance as in Type 1 and 2 models (Figs 4 and 5). Model Types 3 and 4 show that shortening of the accreted terrane crust is inhibited when its lithospheric mantle is depleted or strong, but these cases still exhibit some suture zone advance (Figs 6 and 7).

6.3 Northward underthrusting of the orogen by Indian (pro-)lithosphere

The present-day position of the S-point (the point at which the subducting lithosphere detaches from the crust and descends into the mantle) provides another essential constraint on the lithospheric-scale evolution of the H-T orogen. One possibility (Argand, 1924) is that the S-point moves to the northern limit of the plateau causing its uplift. Modern, less dramatic interpretations indicate that shallowly subducting/flat slab Indian lithospheric mantle underlies the southern plateau. The extent of inferred underthrusting varies along strike (e.g., Replumaz et al. 2013) and between methodologies (see discussion in DeCelles et al., 2002; Nunn et al., 2013), but in general, the available seismic data indicate that high-velocity Indian lithosphere locally extends at least to the Bangong suture beneath the western H-T orogen (BS in Figs 8a, c; Owens and Zandt, 1997; Li et al., 2008; Nunn et al., 2013 and references therein).

Several of the models presented here reproduce this characteristic H-T geometry in which the S-point beneath the plateau lies north of the surface suture as a result of northward underthrusting of the plateau by pro-lithospheric mantle (Fig. 8a and d, SP). The requirement for this style of underthrusting in the models is that flat-slab subduction of the pro-lithosphere is preceded or accompanied by removal of the accreted terrane mantle lithosphere, a behavior that emerges in model Types 1 and 2, where the accreted terrane lithosphere is weak and dense, and in Type 3 where it is shunted.

It is also seen that the retroward advance of the S-point is much less than the total convergence in both the H-T orogen and the models (Figs 8 and 4–7). By implication,

underthrusting alone cannot account for the total length of Indian lithosphere consumed beneath the orogen. The net behavior is advancing subduction; a combination of underthrusting and subduction. India does not merely indent Asia as in thin sheet models (e.g., England and Houseman, 1986).

6.4. Implications for post-collisional magmatism and uplift

Many studies have related H-T uplift and magmatism to removal of the underlying lithospheric mantle. For example, convective removal may have caused uplift sufficient to explain the transition from N-S shortening to E-W extension across the southern plateau (England and Houseman, 1989), as well as broadly coeval, small volume but widespread adakitic and ultrapotassic magmatism within the Lhasa terrane (e.g., Molnar et al., 1993; Turner et al., 1996; Miller et al., 1999; Williams et al., 2001; Chung et al., 2003). For comparison with the models we use the recent interpretation of Chung et al. (2009) (Fig. 8c) based on these ideas. Initially (Fig. 8c; 50 Ma), lower lithosphere of the Lhasa terrane is 'softened' by pre- to syn-collisional Gangdese Arc/Linzizong volcanism, the latter interpreted to result from syn-collisional rollback and breakoff of the subducting Neo-Tethyan slab. Alternatively, the Lhasa terrane lithosphere may have been metasomatised, and 'soft', since the Archaean or Proterozoic (Turner et al., 1996). Subsequent convergence shortens/thickens the Lhasa terrane, which develops a weakened welt at its leading edge (Fig. 8c; 45–30 Ma), and later becomes a Rayleigh-Taylor instability that drips into the mantle (Fig. 8c; 30–25 Ma). Hot sublithospheric mantle upwells in response causing rapid isostatic uplift of the plateau, and leading to the onset of E-W extension. In addition, it heats the base of the thinned lithosphere leading to local melting of variable composition eclogitic lower crust and remaining mantle, resulting in magmatism with adakitic and ultrapotassic affinities. From 20–10 Ma the Lhasa terrane is underthrust at a shallow angle by the subducting Indian lithosphere, leading to the eventual cessation of magmatism at ~10 Ma (Fig. 8c; 20–10 Ma) (Chung et al., 2009).

Type 1 and 2 models presented here produce a similar overall sequence of events (Fig. 8d), with removal of the lithospheric mantle leading to an upwelling of hot sublithospheric mantle and, in some cases, appreciable uplift (Figs 4 and 5). Moreover, the resulting thermal perturbation after delamination (e.g., Moho temperatures of ~1100°C) is consistent with the temperatures required to generate lithospheric and lower crustal melts

of the observed compositions (Turner et al., 1996; Chung et al., 2003). The primary difference lies in the mechanism of lithospheric removal. Whereas Chung et al. (2009) adopt a conceptual Rayleigh-Taylor instability mechanism, our models show that very different styles of lithospheric removal may emerge. In Model AT-1, the accreted terrane 'Lhasa' lithosphere delaminates and sinks from north to south, prior to underthrusting by buoyant 'Indian' pro-lithosphere (Fig. 8d). In contrast, Model AT-2 removes the 'Lhasa' mantle lithosphere by southward entrainment as envisaged conceptually by Molnar (1988). Whether the specific mechanism of mantle lithosphere removal can be diagnosed remains to be determined, perhaps from data on uplift and magmatic sweeps. For example, it is speculated that a north to south magmatic sweep, with associated extension, was caused by rollback of the subducting Indian slab (Kapp et al., 2010; DeCelles et al., 2011). Based on our models we suggest such a trend in the magmatism might equally be the result of progressive north to south delamination or convective entrainment of the Lhasa lithospheric mantle.

In summary, models presented here contrast strongly with the whole lithosphere thin sheet style of orogenesis by coherent viscous flow of crust and mantle lithosphere (e.g., England and Houseman, 1986). Instead, inherited properties of accreted terranes stamp a lasting imprint on deformation that propagates northward, raising the plateau in steps by shortening the crust of the terrane and decoupling its crust from the mantle lithosphere, which then deforms according to its own preferred style. This type of orogenesis is more reminiscent of the conjectural three-phase Tibet growth model proposed by Tapponnier et al. (2001).

7. Conclusions

Continent-continent collisional orogeny often involves prior accretion of terranes to one or both continents, as in the Himalayan-Tibetan orogen. The properties of the accreted terranes can have important effects on the ensuing collision. We have used 2D vertical section finite element thermomechanical models of creeping flows to examine continental collisional orogeny in which a terrane has undergone soft accretion to the retro-lithosphere prior to collision. The models show that the style of accreted terrane lithospheric deformation depends strongly on the density of the subducting pro-continent, as well as

the density and strength of the accreted terrane lithospheric mantle, with the latter linked to the degree to which the lithosphere has been depleted or alternatively metasomatized prior to terminal collision. Models with relatively dense and weak accreted terrane mantle lithospheres produce emergent characteristics, namely delamination of the lithospheric mantle with attendant crustal heating and uplift, that are compatible with current interpretations of the evolution of the H-T orogen. We conclude that delamination of the Lhasa terrane mantle lithosphere, evidenced in the magmatic, structural, and seismic records from the plateau, may have been facilitated by metasomatic densification and weakening.

Acknowledgements

This project was generously supported by the National Science Foundation Continental Dynamics project EAR-1008527; Paul Kapp et al., ‘Collaborative Research: The Suturing Process: Insight from the India-Asia Collision Zone’ through a sub-contract, and by an NSERC Discovery Grant to CB. CB also acknowledges support from the Canada Research Chair in Geodynamics. Numerical models used the software SOPALE-nested developed from SOPALE by D. Guptill, B. Lee and members of the Dalhousie Geodynamics Group. SOPALE was originally developed at Dalhousie University by P. Fullsack. We thank Paul Kapp and members of the ‘Suture’ team, and Becky Jamieson for discussions critical to the success of this research. We also thank Becky Jamieson for comments on an earlier version of the manuscript, Taras Gerya and Alexander Webb for their constructive reviews, and An Yin for editorial comments.

References

- Argand, E. (1924). La Tectonique de l'Asie. Extrait du Compte-rendu du XIIIe Congrès géologique international 1922 (Liège), 1, 171–372.
- Beaumont, C., Jamieson, R.A., Nguyen, M.H. and Lee, B., 2001. Himalayan tectonics explained by extrusion of a low-viscosity crustal channel coupled to focused surface denudation. *Nature*, 414, 738–742.
- Beaumont, C., Jamieson, R.A., Nguyen, M.H. and Medvedev, S., 2004. Crustal channel flows: 1. Numerical models with applications to the tectonics of the Himalayan-Tibetan orogen. *J. Geophysical Research: Solid Earth*, 109, B06406, <http://dx.doi.org/10.1029/2003JB002809>.
- Beaumont, C., Nguyen, M.H., Jamieson, R.A. and Ellis, S., 2006. Crustal flow modes in large hot orogens. In: *Special Publications*, vol. 268. Geological Society, London, pp. 91–145.
- Beaumont, C., Jamieson, R.A., Butler, J.P. and Warren, C.J., 2009. Crustal structure: A key constraint on the mechanism of ultra-high-pressure rock exhumation. *Earth and Planetary Science Letters*, 287, 116–129.
- Billen, M.I., 2010. Slab dynamics in the transition zone. *Physics of the Earth and planetary interiors*, 183, 296–308.
- Butler, J.P., Beaumont, C. and Jamieson, R.A., 2013. The Alps 1: A working geodynamic model for burial and exhumation of (ultra) high-pressure rocks in Alpine-type orogens. *Earth and Planetary Science Letters*, 377, 114–131.
- Butler, J.P., Beaumont, C. and Jamieson, R.A., 2014. The Alps 2: Controls on crustal subduction and (ultra) high-pressure rock exhumation in Alpine-type orogens. *Journal of Geophysical Research: Solid Earth*, 119, 5987–6022.
- Cawood, P.A., Kröner, A., Collins, W.J., Kusky, T.M., Mooney, W.D. and Windley, B.F., 2009. Accretionary orogens through Earth history. Geological Society, London, *Special Publications*, 318, 1–36.
- Chung, S.L., Liu, D., Ji, J., Chu, M.F., Lee, H.Y., Wen, D.J., Lo, C.H., Lee, T.Y., Qian, Q. and Zhang, Q., 2003. Adakites from continental collision zones: melting of thickened lower crust beneath southern Tibet. *Geology*, 31(11), 1021–1024.
- Chung, S.L., Chu, M.F., Ji, J., O'Reilly, S.Y., Pearson, N.J., Liu, D., Lee, T.Y. and Lo, C.H., 2009. The nature and timing of crustal thickening in Southern Tibet: geochemical and zircon Hf isotopic constraints from postcollisional adakites. *Tectonophysics*, 477(1), 36–48.

- Conrad, C.P. and Molnar, P., 1997. The growth of Rayleigh-Taylor-type instabilities in the lithosphere for various rheological and density structures. *Geophysical Journal International*, 129, 95–112.
- DeCelles, P.G., Robinson, D.M. and Zandt, G., 2002. Implications of shortening in the Himalayan fold-thrust belt for uplift of the Tibetan Plateau. *Tectonics*, 21(6), 1062, <http://dx.doi.org/10.1029/2001TC001322>.
- DeCelles, P.G., Kapp, P., Quade, J., and G.E. Gehrels, G.E., 2011. Oligocene–Miocene Kailas basin, southwestern Tibet: Record of postcollisional upper-plate extension in the Indus-Yarlung suture zone. *Geological Society of America Bulletin*, 123, 1337–1362, <http://dx.doi.org/10.1130/B30258.1>
- Dewey, J.F., 1969. Evolution of the Appalachian/Caledonian orogen. *Nature*, 222, 124–129.
- Dewey, J.F., Shackleton, R.M., Chengfa, C. and Yiyin, S., 1988. The tectonic evolution of the Tibetan Plateau. *Philosophical Transactions of the Royal Society of London A: Mathematical, Physical and Engineering Sciences*, 327, 379–413.
- Dupont-Nivet, G., Lippert, P.C., Van Hinsbergen, D.J., Meijers, M.J. and Kapp, P., 2010. Palaeolatitude and age of the Indo–Asia collision: palaeomagnetic constraints. *Geophysical Journal International*, 182, 1189–1198.
- Ellis, S., Beaumont, C., Jamieson, R.A. and Quinlan, G., 1998. Continental collision including a weak zone: the vise model and its application to the Newfoundland Appalachians. *Canadian Journal of Earth Sciences*, 35, 1323–1346.
- Elkins-Tanton, L.T., 2005. Continental magmatism caused by lithospheric delamination. *Geological Society of America Special Papers*, 388, 449–461.
- England, P. and Houseman, G., 1986. Finite strain calculations of continental deformation: 2. Comparison with the India-Asia collision zone. *Journal of Geophysical Research: Solid Earth*, 91(B3), 3664–3676.
- England, P. and Houseman, G., 1989. Extension during continental convergence, with application to the Tibetan Plateau. *Journal of Geophysical Research: Solid Earth*, 94(B12), 17561–17579.
- Fullsack, P., 1995. An arbitrary Lagrangian–Eulerian formulation for creeping flows and applications in tectonic models. *Geophysical Journal International*, 120, 1–23.
- Göğüş, O.H. and Pysklywec, R.N., 2008. Mantle lithosphere delamination driving plateau uplift and synconvergent extension in eastern Anatolia. *Geology*, 36, 723–726.
- Gleason, G.C. and Tullis, J., 1995. A flow law for dislocation creep of quartz aggregates determined with the molten salt cell. *Tectonophysics*, 247, 1–23.

- Griffin, W.L., O'Reilly, S.Y., Afonso, J.C. and Begg, G.C., 2009. The composition and evolution of lithospheric mantle: a re-evaluation and its tectonic implications. *Journal of Petrology*, 50, 1185–1204.
- Hafkenscheid, E., Wortel, M.J.R., Spakman, W., 2006. Subduction history of the Tethyan region derived from seismic tomography and tectonic reconstructions. *Journal of Geophysical Research: Solid Earth*, 111, <http://dx.doi.org/10.1029/2005jb003791>.
- Houseman, G.A., McKenzie, D.P. and Molnar, P., 1981. Convective instability of a thickened boundary layer and its relevance for the thermal evolution of continental convergent belts. *Journal of Geophysical Research: Solid Earth*, 86(B7), 6115–6132.
- Houseman, G.A. and Molnar, P., 1997. Gravitational (Rayleigh–Taylor) instability of a layer with non-linear viscosity and convective thinning of continental lithosphere. *Geophysical Journal International*, 128, 125–150.
- Huang, W., Dupont-Nivet, G., Lippert, P.C., Hinsbergen, D.J., Dekkers, M.J., Waldrup, R., Ganerød, M., Li, X., Guo, Z. and Kapp, P., 2015. What was the Paleogene latitude of the Lhasa terrane? A reassessment of the geochronology and paleomagnetism of Linzizong volcanic rocks (Linzhou basin, Tibet). *Tectonics*, 34, 594–622.
- Johnson, M.R.W., 2002. Shortening budgets and the role of continental subduction during the India–Asia collision. *Earth-Science Reviews*, 59, 101–123.
- Kapp, P.A., Decelles, P.G., Ding, L. and van Hinsbergen, D.J., 2010, December. Cretaceous–Cenozoic Geological Evolution of Tibet: Tectonic Interpretations and Outstanding Questions. In *AGU Fall Meeting Abstracts*, 1.
- Karato, S.I. and Wu, P., 1993. Rheology of the upper mantle- A synthesis. *Science*, 260, 771–778.
- Klemperer, S.L., 2006. Crustal flow in Tibet: geophysical evidence for the physical state of Tibetan lithosphere and inferred patterns of active flow. In: *Special Publications*, vol. 268. Geological Society, London, pp. 39–70.
- Krystopowicz, N.J. and Currie, C.A., 2013. Crustal eclogitization and lithosphere delamination in orogens. *Earth and Planetary Science Letters*, 361, 195–207.
- Lee, C.T.A., 2003. Compositional variation of density and seismic velocities in natural peridotites at STP conditions: Implications for seismic imaging of compositional heterogeneities in the upper mantle. *Journal of Geophysical Research: Solid Earth*, 108(B9).
- Le Pichon, X., Fournier, M. and Jolivet, L., 1992. Kinematics, topography, shortening, and extrusion in the India-Eurasia collision. *Tectonics*, 11, 1085–1098.

- Le Pourhiet, L., Gurnis, M. and Saleeby, J., 2006. Mantle instability beneath the Sierra Nevada mountains in California and Death Valley extension. *Earth and Planetary Science Letters*, 251(1), 104–119.
- Li, C., van der Hilst, R.D., Meltzer, A.S., Engdahl, E.R., 2008. Subduction of the Indian lithosphere beneath the Tibetan Plateau and Burma. *Earth and Planetary Science Letters*, 274, <http://dx.doi.org/10.1016/j.epsl.2008.07.016>.
- Liang, X., Sandvol, E., Chen, Y.J., Hearn, T., Ni, J., Klemperer, S., Shen, Y. and Tilmann, F., 2012. A complex Tibetan upper mantle: A fragmented Indian slab and no south-verging subduction of Eurasian lithosphere. *Earth and Planetary Science Letters*, 333, 101–111.
- Lippert, P.C., Van Hinsbergen, D.J. and Dupont-Nivet, G., 2014. Early Cretaceous to present latitude of the central proto-Tibetan Plateau: A paleomagnetic synthesis with implications for Cenozoic tectonics, paleogeography, and climate of Asia. *Geological Society of America Special Papers*, 507, SPE507-01.
- Mackwell, S.J., Zimmerman, M.E. and Kohlstedt, D.L., 1998. High-temperature deformation of dry diabase with application to tectonics on Venus. *Journal of Geophysical Research: Solid Earth*, 103, 975–984.
- Miller, C., Schuster, R., Klötzli, U., Frank, W. and Purtscheller, F., 1999. Post-collisional potassic and ultrapotassic magmatism in SW Tibet: geochemical and Sr–Nd–Pb–O isotopic constraints for mantle source characteristics and petrogenesis. *Journal of Petrology*, 40(9), 1399–1424.
- Molnar, P., 1988. A Review of Geophysical Constraints on the Deep Structure of the Tibetan Plateau, the Himalaya and the Karakoram, and their Tectonic Implications, *Phil. Trans. R. Soc.: Math. Phys. Eng. Sci.*, 326(1589), 33–88.
- Molnar, P., Houseman, G.A. and Conrad, C.P., 1998. Rayleigh-Taylor instability and convective thinning of mechanically thickened lithosphere: effects of non-linear viscosity decreasing exponentially with depth and of horizontal shortening of the layer. *Geophysical Journal International*, 133, 568–584.
- Molnar, P., England, P. and Martinod, J., 1993. Mantle dynamics, uplift of the Tibetan Plateau, and the Indian monsoon. *Reviews of Geophysics*, 31, 357–396.
- Morency, C. and Doin, M.-P., 2004. Numerical simulations of the mantle lithosphere delamination. *Journal of Geophysical Research: Solid Earth*, 109, B03410, <http://dx.doi.org/10.1029/2003JB002414>.
- Moresi, L., Betts, P.G., Miller, M.S., and Cayley, R.A., 2014. Dynamics of continental accretion. *Nature*, 508, 245–248, <http://dx.doi.org/10.1038/nature13033>.
- Nábělek, J., Hetényi, G., Vergne, J., Sapkota, S., Kafle, B., Jiang, Mei, Heping, Su., Chen,

- John, Huang, Bor-Shouh, the Hi-CLIMB Team, 2009. Underplating in the Himalaya–Tibet collision zone revealed by the Hi-CLIMB experiment. *Science* 325. <http://dx.doi.org/10.1126/science.1167719>.
- Nunn, C., Roecker, S.W., Tilmann, F.J., Priestley, K.F., Heyburn, R., Sandvol, E.A., Ni, J.F., Chen, Y.J., Zhao, W. and INDEPTH Team, 2013. Imaging the lithosphere beneath NE Tibet: teleseismic P and S body wave tomography incorporating surface wave starting models. *Geophysical Journal International*, <http://dx.doi.org/10.1093/gji/ggt476>.
- Nunn, C., Roecker, S.W., Priestley, K.F., Liang, X. and Gilligan, A., 2014. Joint inversion of surface waves and teleseismic body waves across the Tibetan collision zone: the fate of subducted Indian lithosphere. *Geophysical Journal International*, 198, 1526–1542.
- O'Reilly S., and Griffin, W.L., 2013. Mantle Metasomatism, Chapter 12 in D.E. Harlov and H. Austrheim, *Metasomatism and the Chemical Transformation of Rock*, Lecture Notes in Earth System Sciences, http://dx.doi.org/10.1007/978-3-642-28394-9_12, Springer-Verlag Berlin Heidelberg.
- Owens, T.J. and Zandt, G., 1997. Implications of crustal property variations for models of Tibetan plateau evolution. *Nature*, 387, 37–43.
- Replumaz, A., Karason, H., van der Hilst, R.D., Besse, J., Tapponnier, P., 2004. 4-D evolution of SE Asia's mantle from geological reconstructions and seismic tomography. *Earth and Planetary Science Letters*, 221, [http://dx.doi.org/10.1016/s0012-821x\(04\)00070-6](http://dx.doi.org/10.1016/s0012-821x(04)00070-6).
- Replumaz, A., Negredo, A.M., Guillot, S., Villaseñor, A., 2010a. Multiple episodes of continental subduction during India/Asia convergence: insight from seismic tomography and tectonic reconstruction. *Tectonophysics*, <http://dx.doi.org/10.1016/j.tecto.2009.10.007>.
- Replumaz, A., Negredo, A.M., Guillot, S., van der Beek, P., Villaseñor, A., 2010b. Crustal mass budget and recycling during the India/Asia collision. *Tectonophysics*, <http://dx.doi.org/10.1016/j.tecto.2009.10.007>.
- Replumaz, A., Negredo, A.M., Villaseñor, A., Guillot, S., 2010c. Indian continental subduction and slab break-off during Tertiary Collision. *Terra Nova*, <http://dx.doi.org/10.1111/j.1365-3121.2010.00945.x>.
- Replumaz, A., Guillot, S., Villaseñor, A. and Negredo, A.M., 2013. Amount of Asian lithospheric mantle subducted during the India/Asia collision. *Gondwana Research*, 24(3), 936–945.
- Roberts, D., 2003. The Scandinavian Caledonides: event chronology, palaeogeographic settings and likely modern analogues. *Tectonophysics*, 365(1), 283–299.

902 Rosenberg, C.L. and Handy, M.R., 2005. Experimental deformation of partially melted
903 granite revisited: implications for the continental crust. *Journal of Metamorphic Geology*,
904 23, 19–28.

905
906 Royden, L.H., Burchfiel, B.C. and van der Hilst, R.D., 2008. The geological evolution of the
907 Tibetan Plateau. *Science*, 321, 1054–1058.

908
909 Shen, F., Royden, L.H. and Burchfiel, B.C., 2001. Large-scale crustal deformation of the
910 Tibetan Plateau. *Journal of Geophysical Research*, 106, 6793–6816.

911
912 Sobolev, S.V. and Babeyko, A.Y., 2005. What drives orogeny in the Andes? *Geology*, 33(8),
913 617–620.

914
915 Sternai, P., Avouac, J.-P., Jolivet, L., Faccenna, C., Gerya, T., Becker, T.W., Menant, A., 2016. On
916 the influence of the asthenospheric flow on the tectonics and topography at a collision-
917 subduction transition zones: Comparison with the eastern Tibetan margin. *Journal of*
918 *Geodynamics*, <http://dx.doi.org/10.1016/j.jog.2016.02.009>.

919 Tapponnier, P., Zhiqin, X., Roger, F., Meyer, B., Arnaud, N., Wittlinger, G. and Jingsui, Y.,
920 2001. Oblique stepwise rise and growth of the Tibet Plateau. *Science*, 294, 1671–1677.

921
922 Tetreault, J.L. and Buitier, S.J.H., 2014. Future accreted terranes: a compilation of island arcs,
923 oceanic plateaus, submarine ridges, seamounts, and continental fragments. *Journal of*
924 *Geophysical Research: Solid Earth*, 5, 1243–1275.

925
926 Turner, S., Arnaud, N., LIU, J., Rogers, N., Hawkesworth, C., Harris, N., Kelley, S., Van
927 Calsteren, P. and Deng, W., 1996. Post-collision, shoshonitic volcanism on the Tibetan
928 Plateau: implications for convective thinning of the lithosphere and the source of ocean
929 island basalts. *Journal of petrology*, 37(1), 45–71.

930
931 Van der Voo, R., Spakman, W. and Bijwaard, H., 1999. Tethyan subducted slabs under India.
932 *Earth and Planetary Science Letters*, 171, 7–20.

933
934 van Hinsbergen, D.J.J., P. Kapp, G. Dupont-Nivet, P.C. Lippert, P.G. DeCelles, and T.H. Torsvik,
935 2011a. Restoration of Cenozoic deformation in Asia and the size of Greater India, *Tectonics*,
936 30, TC5003, <http://dx.doi.org/10.1029/2011TC002908>.

937
938 van Hinsbergen, D.J.J., B. Steinberger, P. V. Doubrovine, and R. Gassmüller, 2011b.
939 Acceleration and deceleration of India-Asia convergence since the Cretaceous: Roles of
940 mantle plumes and continental collision, *J. Geophys. Res.*, 116, B06101,
941 <http://dx.doi.org/10.1029/2010JB008051>.

942
943 Van Staal, C.R., Dewey, J.F., Mac Niocaill, C. and McKerrow, W.S., 1998. The Cambrian-
944 Silurian tectonic evolution of the northern Appalachians and British Caledonides: history of
945 a complex, west and southwest Pacific-type segment of Iapetus. *Geological Society, London*,
946 *Special Publications*, 143, 197–242.

- 947
948 Vogt, K. and Gerya, T.V., 2014. From oceanic plateaus to allochthonous terranes: Numerical
949 modelling. *Gondwana Research*, 25, 494–508.
950
- 951 Warren, C. J., C. Beaumont, and R. A. Jamieson, 2008. Deep subduction and rapid
952 exhumation: Role of crustal strength and strain weakening in continental subduction and
953 ultrahigh-pressure rock exhumation, *Tectonics*, 27, TC6002,
954 <http://dx.doi.org/10.1029/2008TC002292>.
955
- 956 Willett, S.D. and Beaumont, C., 1994. Subduction of Asian lithospheric mantle beneath Tibet
957 inferred from models of continental collision. *Nature*, 369, 642–645.
958
- 959 Williams, H. and Hatcher, R.D., 1982. Suspect terranes and accretionary history of the
960 Appalachian orogen. *Geology*, 10, 530–536.
961
- 962 Williams, H., Turner, S., Kelley, S. and Harris, N., 2001. Age and composition of dikes in
963 Southern Tibet: New constraints on the timing of east-west extension and its relationship
964 to postcollisional volcanism. *Geology*, 29(4), 339–342.
965
- 966 Wilson, J.T., 1966. Did the Atlantic close and then re-open? *Nature*, 211, 676–681.
967
- 968 Yin, A. and Harrison, T.M., 2000. Geologic evolution of the Himalayan-Tibetan orogen.
969 *Annual Review of Earth and Planetary Sciences*, 28, 211–280
970

Figure and table captions

Figure 1: (a) Design of two-dimensional models (see Section 2 and Appendix 1 for details). The full model 4000×1200 km domain is represented by a coarse computational grid with a maximum resolution of 10×2 km. A higher-resolution, ‘nested’ computational grid (dashed line) is embedded within the coarse mesh in the vicinity of the subduction zone and retro-lithosphere, and has a maximum resolution of 2×2 km. AT = accreted terrane, CLM = continental lithospheric mantle, OLM = oceanic lithospheric mantle, SLM = sublithospheric mantle. Black arrows show velocity boundary conditions in schematic form. Red shows thermal boundary conditions (basal heat flux) and resulting surface heat flux and continental temperatures. AT = accreted terrane, P = pro-continent lithosphere, R = retro-continent lithosphere, OLM = oceanic lithospheric mantle, SLM = sublithospheric mantle. WQ = wet quartzite, DMD = dry Maryland diabase, WOL = wet olivine. ϕ = range of effective angle of internal friction (ϕ_{eff}) owing to pore-fluid pressure and frictional-plastic strain-softening; A = radiogenic heat production ($\mu\text{W m}^{-3}$) (A_R in text); ρ = density (Table 1). (b) Schematic illustration of experimental parameter space (see Section 4). Panels in 3-dimensional grid correspond to Figures 3a–c. Each panel represents a set of models with a different pro-continent CLM density. Within each panel, models vary by AT-CLM density and AT-CLM strength (viscosity scaling factor f).

Figure 2: Results for Model AT-PR, P($\Delta\rho = 25:f = 10$)-AT($\Delta\rho = 50:f = 5$) (prototype model). (a–d) Material colors (see Fig. 1) and isotherms (100°C and then at 200°C increments) for selected model times chosen to show significant model developments for this and subsequent models. Note the overall lack of deformation of the accreted terrane lithosphere – a consequence of its high strength and depletion density. Δx is the post-collisional convergence; red vertical arrow marks position of surface suture at collision, with S marking its evolving position and SP marking the surface above the S-point. MW denotes ‘Melt Weakened’ upper continental crust.

Figure 3: Summary of results of sensitivity experiments. Each set (a, b, and c) corresponds to a different pro-continental lithospheric mantle depletion, 30, 40 and 50 kg m⁻³. Within each set, model experiments vary the accreted terrane lithospheric mantle depletion ($\Delta\rho = 0, 10, 20, 40$ kg m⁻³) and viscosity scaling factor ($f = 2, 3.5, 5$). (a) Experiment set with pro-continent depletion of 30 kg m⁻³. Times are given in millions of years post-collision (Myr-pc). Models are color-coded according to the main model types described in the text. Green = Type 1; Red = Type 2; Purple = Type 3; Blue = Type 4. (b) Experiment set with pro-continent depletion of 40 kg m⁻³. Color coding as in (a). (c; next page) Experiment set with pro-continent depletion of 50 kg m⁻³. Type 1 (green) style models tend to cluster where the pro-continent depletion is high ($\Delta\rho = 50$ kg m⁻³) (panel c), and the accreted terrane mantle lithosphere is viscously weak (mostly $f = 2$ –3.5) and/or has a low depletion density ($\Delta\rho = 0$ –20 kg m⁻³). They exhibit a systematic delay in the onset of delamination as the depletion density of the accreted terrane lithospheric mantle increases; increasing buoyancy (panel c; $f = 2$ column). Type 2 (red) dominates, appearing across a wide range of the parameter space explored here. As in Type 1 models, there is a systematic delay in the timing of delamination as the buoyancy and strength of the accreted terrane lithospheric mantle increase (e.g., panel a). Type 3 (purple) behavior is limited to four models characterized

either by high buoyancy or strength of the accreted terrane mantle lithosphere. Type 4 (blue) models cluster where the accreted terrane lithospheric mantle is strongly depleted, buoyant ($\Delta\rho = 40 \text{ kg m}^{-3}$) and strong ($f = 3.5\text{--}5$), allowing it to buttress the compressive force exerted by the pro-continent effectively.

Figure 4: Results for Model AT-1, P(50:10)-AT(0:2), showing delamination of accreted terrane lithospheric mantle by northward subduction followed by flat-slab underthrusting. (a–d) Material colors and isotherms for selected model times. Isotherms are plotted every 200°C starting at 100°C; for clarity only 700°C (approximate starting Moho temperature) and 1300°C (approximate temperature at base of lithospheric mantle) are labeled. Myr-pc is millions of years post-collision. Δx is the amount of post-collisional convergence; dark red vertical arrow marks position of surface suture at collision, with S marking its evolving position and SP marking the position of the S-point. Inset in panel c shows evolution of model surface elevation during delamination of the accreted terrane lithospheric mantle. MW denotes ‘Melt Weakened’ upper continental crust.

Figure 5: Results for Model AT-2, P(40:10)-AT(0:2), showing delamination of accreted terrane lithospheric mantle by entrainment by pro-lithosphere and southward subduction, followed by flat-slab underthrusting. (a–d) Material colors and isotherms for selected model times. Isotherms are plotted every 200°C starting at 100°C; for clarity only 700°C (approximate starting Moho temperature) and 1300°C (approximate temperature at base of lithospheric mantle) are labeled. Myr-pc is millions of years post-collision. Δx is the amount of post-collisional convergence; dark red vertical arrow marks position of surface suture at collision, with S marking its evolving position and SP marking the position of the S-point. Inset in panel c shows evolution of model surface elevation during delamination of the accreted terrane lithospheric mantle. MW denotes ‘Melt Weakened’ upper continental crust.

Figure 6: Results for Model AT-3, P(50:10)-AT(40:2), showing northward shunting of the accreted terrane lithospheric mantle. (a–d) Material colors and isotherms for selected model times. Isotherms are plotted every 200°C starting at 100°C; for clarity only 700°C (approximate starting Moho temperature) and 1300°C (approximate temperature at base of lithospheric mantle) are labeled. Myr-pc is millions of years post-collision. Δx is the amount of post-collisional convergence; red vertical arrow marks position of surface suture at collision, with S marking its evolving position and SP marking the position of the S-point. Inset in panel c shows evolution of model surface elevation during shunting of the accreted terrane lithospheric mantle. MW denotes ‘Melt Weakened’ upper continental crust.

Figure 7: Results for Model AT-4, P(40:10)-AT(20:5), showing smearing style deformation of the accreted terrane lithospheric mantle. (a–d) Material colors and isotherms for selected model times. Isotherms are plotted every 200°C starting at 100°C; for clarity only 700°C (approximate starting Moho temperature) and 1300°C (approximate temperature at base of lithospheric mantle) are labeled. Δx is the amount of post-collisional convergence; red vertical arrow marks position of surface suture at collision, with S marking its evolving position and SP marking the position of the S-point. Myr-pc is millions of years post-collision. Inset in panel d shows evolution of model surface elevation during smearing of

the accreted terrane lithospheric mantle. MW denotes ‘Melt Weakened’ upper continental crust.

Figure 8: Summary of first-order constraints on the evolution (shown from bottom to top) of the Himalayan-Tibetan orogen (left column) and comparison with model results (right column). (a) Schematic interpretation of current lithospheric-scale structure of the H-T orogen derived from seismic data, modified from Owens and Zandt (1997) (upper panel) and Nunn et al. (2014) (lower panel). Bold labels correspond to major structure/terrane boundaries; ITS = Indus-Tsangpo Suture; BS=Bangong Suture; JS = Jinsha Suture; KS = Kunlun Fault/Suture; S-G = Songpan-Ganzi terrane MCT = Main Central Thrust; STD = South Tibetan Detachment. Note the inferred underthrusting of the Indian mantle lithosphere to beneath the Bangong Suture and further north. (b) Inferred mantle structure beneath the H-T system based on seismic tomography, showing remnant subducted slabs, modified from Van der Voo et al. (1999). Roman numerals denote seismic anomalies interpreted as subducted slabs derived from closure of Mesotethys (I) (which separated the Qiangtang and Lhasa terranes), Neotethys (II and III), and subducted Indian mantle lithosphere (IV). (c) Schematic interpretation of the geodynamical evolution responsible for post-collisional magmatism in the Lhasa terrane, modified from Chung et al. (2009). (d) Geologic interpretation of Model AT-1 (AT-2 orange stipple) in the context of the H-T system. ‘India’, ‘Lhasa’ and ‘Qiangtang’ highlight parts of the model corresponding broadly to elements within the H-T system. For alternative interpretations of the lithospheric-scale structure of the H-T system, see Klemperer (2006), Li et al. (2008), Hafkenscheid et al. (2006), Nábělek et al. (2009), Replumaz et al. (2004; 2010a; 2010b; 2010c), and Liang et al. (2012).

Table 1:

Material and geometrical parameters used in model experiments. AT = accreted terrane, P = pro-continent, R = retro-continent. Flow laws are dry Maryland diabase (DMD; Mackwell et al., 1998), wet olivine (WOL; Karato and Wu, 1993), and wet quartzite (WQ; Gleason and Tullis, 1995).

Figure Captions for supplementary material

Figure S1: Results for Model AT-2, P(40:10)-AT(0:2), showing delamination of accreted terrane lithospheric mantle by entrainment by pro-lithosphere and southward subduction, followed by flat-slab underthrusting. (a–d) Material colors and isotherms for selected model times. Isotherms are plotted every 200°C starting at 100°C; for clarity only 700°C (approximate starting Moho temperature) and 1300°C (approximate temperature at base of lithospheric mantle) are labeled. Myr-pc is millions of years post-collision. Δx is the amount of post-collisional convergence; dark red vertical arrow marks position of surface suture at collision, with S marking its evolving position and SP marking the position of the S-point. MW denotes ‘Melt Weakened’ upper continental crust.

Figure S2: Results for Model AT-2, P(40:10)-AT(0:2) (see Figure S1) showing temperature-dependent density of model materials. A restricted density range (3100–3300 kg m⁻³) is shown in order to highlight the density contrast between the subducting pro-lithospheric

1106 slab and the sublithospheric mantle. Part of the subducted pro-lithospheric slab in the
1107 model is positively buoyant owing to its temperature and relative density depletion of 40
1108 kg m^{-3} with respect to the sublithospheric mantle. Isotherms are plotted every 200°C
1109 starting at 100°C; for clarity only 700°C (approximate starting Moho temperature) and
1110 1300°C (approximate temperature at base of lithospheric mantle) are labeled. Myr-pc is
1111 millions of years post-collision. Δx is the amount of post-collisional convergence.

1112 **Figure S3:** Results for Model AT-2, P(40:10)-AT(0:2) (see Figure S1) showing the viscosity
1113 of the model materials. Isotherms are plotted every 200°C starting at 100°C; for clarity only
1114 700°C (approximate starting Moho temperature) and 1300°C (approximate temperature at
1115 base of lithospheric mantle) are labeled. Myr-pc is millions of years post-collision. Δx is the
1116 amount of post-collisional convergence. Note the regions of low viscosity, melt weakened,
1117 continental upper crust where temperature exceeds 700°C.

Appendices

A1. Modeling Methods

A1.1 Overview of SOPALE Nested code

We use two-dimensional thermal-mechanical models computed using the software SOPALE Nested (Fallsack, 1995; Beaumont et al., 2009; Butler et al 2014) to investigate the dynamics of collisional orogens at the upper-mantle and part lower-mantle-scale. The models are designed based on a simple geometry representing oceanic subduction followed by continent-continent collision in which one continent contains an already accreted terrane (Figure 1). The methodology remains similar to our recent work (Beaumont et al., 2009; Butler et al., 2014).

The models are computed by solving the equations for incompressible creeping (Stokes) flows (Eqs. A.1 and A.2) and energy balance (Eq. A.3) subject to mechanical and thermal boundary conditions:

$$\text{(Eq. A.1)} \quad \frac{\partial \sigma_{ij}}{\partial x_i} - \frac{\partial P}{\partial x_i} + \rho g = 0, \quad i, j = 1, 2$$

$$\text{(Eq. A.2)} \quad \frac{\partial v_i}{\partial x_i} = 0, \quad i = 1, 2$$

$$\text{(Eq. A.3)} \quad \rho(T) c_p \left(\frac{\partial T}{\partial t} + v_i \frac{\partial T}{\partial x_i} \right) = \frac{\partial}{\partial x_i} K(T) \frac{\partial T}{\partial x_i} + A_R + A_{SH} + v_2 \alpha g T \rho(T), \quad i = 1, 2,$$

where σ_{ij} is the deviatoric stress tensor, x_i spatial coordinates, P pressure (mean stress), ρ density, g gravitational acceleration, v_i a component of velocity, c_p specific heat, T temperature, t time, K thermal conductivity, A_R (crustal) radiogenic heat production per unit volume, A_{SH} shear heating, and α volumetric thermal expansion. Coupling of the mechanical and thermal solutions is achieved through the advection of radioactive crust, shear heating, thermal activation of viscous flow, and buoyancy forces arising from metamorphic phase changes and thermal expansion.

The models use a sub-grid approach in which a higher-resolution computational domain is ‘nested’ within a larger, lower-resolution domain. The velocity and thermal solution is first obtained for the full domain. This solution is then interpolated onto the boundaries of the nested domain encompassing the vicinity of the subduction zone and the accreted terrane. Coupling of the grids is maintained by using a single cloud of Lagrangian

particles, which is passed between the two solutions, with the particles within the nested domain always obeying the higher resolution solution.

A1.2 Model geometry and boundary conditions

The model domain is 4000 km wide by 1200 km deep. The initial configuration of the prototype model is shown in Figure 1. The starting geometry of the models (Fig. 1a; Table 1) represents a simplified continent-continent collisional system that comprises (left to right) a pro-continent, a 1000-km-wide ocean, and a composite retro-lithosphere, overlying upper mantle to depth of 660 km and then lower mantle to 1200 km. The retro-lithosphere contains a 900-km-wide accreted terrane, and an interior, older stronger retro-continent. The pro-continent upper crust is 24 km thick. The accreted terrane and retro-continent upper crusts are 20 km thick and are overlain by a 4-km-thick sedimentary cover layer. The continental lower crust is everywhere 12 km thick, and the continental lithospheric mantle is 84 km thick (i.e., extending to a depth of 120 km). The oceanic crust is 8 km thick with an overlying 2 km thick sedimentary cover layer. The oceanic lithospheric mantle is 90 km thick, with the upper 50 km chemically depleted and less dense. The lower mantle/upper mantle boundary is at 660 km. Other geometrical properties are given in Table 1.

Subduction is initiated by a weak seed embedded between ocean and accreted terrane that spans both the accreted terrane crust and lithospheric mantle. The boundary conditions applied to the models comprise a pro-lithospheric convergence velocity $V_p = 15$ and 5 cm a^{-1} (oceanic subduction and continental collision, respectively) on the left lithospheric boundary and a retro-lithospheric velocity, $V_r = 0 \text{ cm a}^{-1}$ on the right boundary. Small sublithospheric boundary fluxes are used to ‘pump’ material out of the sides of the model to conserve volume and maintain isostatic equilibrium. The sides of the domain have vertical free slip boundary conditions and the base has no motion boundary conditions. The top of the model is a free surface. Thermal boundary conditions include a basal heat flux of 21 mW m^{-2} , a 0°C surface, and insulated side boundaries. The initial temperature regime corresponds to 2D thermal steady state.

A1.3 Material properties

The flow laws and material properties are similar to those used in Beaumont et al. (2001), and Butler et al. (2013). For a more detailed justification for the choice of flow laws and the scaling of these flow laws see Butler et al. (2014). As mentioned above

the parameters of the prototype model are considered typical of continental and oceanic lithospheric materials (Beaumont et al., 2006, and refs. therein).

Model materials deform by either frictional-plastic (brittle) or viscous (ductile) flow that includes additional strain-softening/strain-weakening mechanisms. Brittle deformation is modeled using the frictional-plastic Drucker-Prager yield criterion, σ_y :

$$(Eq. A.4) \quad \sigma_y = (J'_2)^{1/2} = P \sin \phi_{\text{eff}} + C \cos \phi_{\text{eff}}$$

where J'_2 is the second invariant of the deviatoric stress, P dynamical pressure (mean stress), and C cohesion. The effective angle of internal friction, ϕ_{eff} , is defined to include the effects of pore fluid pressure. Ductile deformation is modeled using power-law flow, for which the effective viscosity is:

$$(Eq. A.5) \quad \eta_{\text{eff}}^v = \frac{f}{W_s} A^{-1/n} \dot{I}'_2^{(1-n/2n)} \exp\left(\frac{Q+PV^*}{nRT_K}\right)$$

where f is a viscosity scaling factor, W_s a strain-weakening factor, A the pre-exponential factor, converted to the tensor invariant form, $\dot{I}'_2$ the second invariant of the deviatoric strain rate, n the stress exponent, Q activation energy, P pressure, V^* the activation volume for power-law creep, T_K absolute temperature, and R the universal gas constant. For simplicity, we base the model materials on a small set of laboratory-determined flow laws.

If the deviatoric stress in the mantle lithosphere reaches a high stress limit, the Peierls stress, σ_p , a new rheology dominates where the viscosity has an exponential dependence on stress. SOPALE models this rheological law by effectively limiting the calculated stress at, σ_p , by downscaling the viscosity so that the stress does not exceed the Peierls stress (100 MPa and 350 MPa) for oceanic ('wet') and continental ('dry') lithospheric mantle, respectively).

Crustal materials and the lithospheric mantle soften/weaken in the frictional-plastic and viscous regimes, respectively. Frictional-plastic materials strain-soften through a linear decrease in their effective angle of internal friction, ϕ_{eff} , with accumulated strain: $\phi_{\text{eff}} =$

$$\phi_{\text{eff}}(\varepsilon). \quad \varepsilon = I'^{1/2}_2, \text{ the square root of the second invariant of deviatoric strain, with } \varepsilon \text{ (strain)}$$

used for simplicity. Viscous strain-weakening occurs through a linear decrease in effective viscosity by the factor W_s over a specified ε range. For all models, crustal materials weaken by $W_s = 10$ over the range $\varepsilon = 5$ –10. Strain-weakening (and softening) as implemented in

our models depends on the resolution of the computational grid (2×2 km in the nested region). Given that natural shear zones form at finer scales, equivalent natural strain weakening/softening occurs at smaller offsets across shear zones because the shear zones are narrower than the typical 6 km in these models. It follows that the models weaken/soften more slowly than ones would with higher resolution. By implication, higher resolution models will give equivalent results with proportionally larger values of the minimum ϕ_{eff} , and smaller values of W_s . The values used here should therefore be interpreted as minimum and maximum values, respectively.

Each crustal unit comprises an upper/middle crust with ϕ_{eff} (hereafter simply ϕ) = 15–2° (except where noted), and a scaled wet quartzite (WQ× f) flow law (Gleason and Tullis, 1995), overlying a lower crust with ϕ = 15–2° (where ϕ = 2° is fully strain-softened) and a dry Maryland diabase (DMD×0.1) flow law, scaled to represent intermediate granulite (Mackwell et al., 1998). The pro-continent and accreted terrane have WQ×1 upper crusts, and the retro-continent a stronger WQ×5 upper crust. The underlying lithospheric mantle extends to 120 km beneath the continents and to 90 km beneath the oceanic plate, has ϕ = 15–2° and a scaled wet olivine flow law (WOL; Karato and Wu, 1993) scaled to represent a somewhat depleted/dehydrated lithosphere, and overlies sublithospheric upper mantle (WOL×2) that extends to 660 km. For the pro-continent and retro-continent, the lithospheric mantle has WOL×10. The accreted terrane lithospheric mantle has a WOL×5 lithospheric mantle in the prototype model, and f = 2, 3.5, and 5 in the sensitivity experiments. Owing to considerable uncertainty in the rheology of the lower mantle (Billen, 2010), we assume that the lower mantle has a constant/uniform viscosity of 1×10^{21} Pa s. The 660-km-deep upper to lower mantle boundary corresponds to a dynamical phase change boundary across which density increases by ~8 %, and the viscosity changes from power-law creep to constant linear viscosity.

The sediment and upper continental crust undergo melt weakening in the same manner as in Beaumont et al. (2001; 2004) such that the viscosity decreases linearly from the calculated power-law value to a fixed linear value of 10^{19} Pa s over the temperature range 700–750°C. This is designed to represent the viscosity reduction that results from

the presence of minor partial melting with a melt fraction of $\sim 4\text{--}7\%$ (Rosenberg and Handy, 2005).

The continents have crustal radiogenic heat production (A in Fig. 1a) for the upper and lower continental crust of 2.0 and $0.4 \mu\text{Wm}^{-3}$, respectively. Coupled with the thermal boundary conditions, the resulting steady state surface heat flow and Moho temperatures on the continents are 62 mWm^{-2} and $\sim 670^\circ\text{C}$. The base lithosphere temperature throughout the model is $\sim 1350^\circ\text{C}$.

A1.4 Density, volume and mass conservation during phase transitions

Crustal materials undergo pressure- and temperature-dependent density changes corresponding to the eclogite and coesite-eclogite phase transitions following Warren et al. (2008) (Table 1), and reverse during exhumation. Mantle materials undergo a similar process involving an $\sim 8\%$ density increase at 660 km depth. During these phase changes the incompressibility equation is modified to that of mass conservation: $\partial\rho/\partial t = -\partial(\rho v_i)/\partial x_i$. This accounts for the volume change and its effect on the buoyancy and velocity field. The volume change is calculated numerically by applying additional normal, compressive/dilatational forces to finite elements at the time they are subject to phase-related density changes. The value of the excess pressure is $\Delta P = \Delta\rho/\beta_v\rho$, where β_v is the viscous bulk modulus of the material, and $\Delta\rho/\rho$ is the fractional change in density corresponding to the phase change. The excess pressure compresses material locally and only during the model time steps when the phase changes occur, thereby ensuring mass conservation. The fractional volume change accompanying a phase change is small in these models and its effect on the velocity field is minor because it only applies at the time of the phase change. However, failure to ensure mass conservation would have a long-term effect on the model because the buoyancy forces will be over- or under-estimated by the fractional error in the material volume.

Appendix A2: Derivation of the Archimedes Number

The selection of the model experimental parameters can be guided by examining the governing equations in non-dimensional form. Equations A.1 and A.2 can be put into non-dimensional form. Consider, for example, a portion of lithospheric mantle, viscosity η_{eff} , with a characteristic length L , embedded in less dense mantle and using the boundary

velocity, V . By scaling distance $x_i^* = x_i/L$, and $dx_i^* = dx_i/L$, velocity $v_i^* = v_i/V$, and pressure $P^* = PL/\eta_{\text{eff}}V$, and where $\rho = \rho_l - \rho_f$ is the difference in densities of the entrained body (lithosphere) and the surrounding fluid, Equations A.1 and A.2 can be rewritten as:

$$\text{(Eq. A.6)} \quad \frac{\partial^2 v_i^*}{\partial x_i^{*2}} - \frac{\partial P^*}{\partial x_j^*} = \frac{g_2 \rho L^2}{V \eta_{\text{eff}}} = \frac{g_2 \rho L^3}{V \eta_{\text{eff}} L} = Ar^{1/2}$$

where g_2 signifies that gravity acts vertically and Ar is the Archimedes number. The numerator of this expression is proportional to the gravitational force on the body, while the denominator is proportional to the viscous drag (e.g., viscous drag on a sphere of radius L). For large values of Ar , corresponding to large contrasts in density ($\rho_l - \rho_f$) and small values of viscosity η_{eff} , the solution is dominated by gravitational forces, while for small values of Ar , the fluid flow is dominated by viscous forces. The Archimedes scaling indicates that in regard to lithospheric mantle processes for a constant V , the main governing model properties are the effective density, that is the density difference between lithospheric mantle and the sublithospheric mantle, and the effective viscosity of the mantle lithosphere, which we take to vary by the factor f , as in Equation A.5. We therefore choose these properties for a sensitivity analysis of model behaviour and interpret the variations in these quantities to reflect among other things the effects of metasomatism on the accreted terrane lithospheric mantle.

Figure 1

Fig. 1

(a) Model design

WW = regions of upper continental crust that have been Melt Weakened ($T > 700^\circ\text{C}$)

(b) Experiment template

Figure 2

Fig. 2

(a) Pro-continent depletion of 30 kg m^{-3} (b) Pro-continent depletion of 40 kg m^{-3}

(c) Pro-continent depletion of 50 kg m^{-3}

Figure 4

Fig. 4

Figure 5

Fig. 5

Figure 6

Fig. 6

Figure 7

Fig. 7

(a) Present-day lithospheric structure

(b) Tomographic interpretation

(c) Tectonic interpretation

(d) Model AT-1 interpretation

Table 1

	Units	Cover (oceanic)	Cover (AT and RC)	PC upper crust	AT upper crust	RC upper crust	Lower crust (all)	Oceanic crust	PC lithospheric mantle	AT lithospheric mantle	RC lithospheric mantle	Oceanic lithospheric mantle	Sub- lithospheric upper mantle	Lower mantle
Mechanical parameters														
Thickness (max.)	km	2	4	24	20	20	12	8	84	84	84	90	to 660 km depth	660–1200 km depth
Reference density (0°C)	kg m ⁻³	2800	2800	2800	2800	2800	2900	2950	3345	3320	3340	3370	3370	3630
Reference density HP (0°C) ^a	kg m ⁻³	2900	2900	2900	2900	No Change	3200	3400	No change	No change	No change	No change	No change	No change
Reference density UHP (0°C) ^a	kg m ⁻³	3000	3000	3000	3000	No Change	No Change	No Change	No change	No change	No change	Changes to 3630 at 660 km depth	Changes to lower mantle at 660 km depth	No change
Effective angle of internal friction (ϕ_{eff}) ^b	deg	8–2	8–2	15–2	15–2	15–2	15–2	15–2	15–2	15–2	15–2	15–2	15–2	15–2
Cohesion	MPa	10	10	2	2	2	2	0	0	0	0	0	0	0
Flow law	–	WQ	WQ	WQ	WQ	WQ	DMD	DMD	WOL	WOL	WOL	WOL	WOL	Constant viscosity 1 × 10 ²¹ Pa s
Viscosity scaling factor (f)	–	1	1	1	1	5	0.1	0.1	10	5	10	2	2	–
Strain weakening factor (W_s)	–	10	10	10	10	10	10	10	1	1	1	1	1	–
n	–	4	4	4	4	4	4.7	4.7	3	3	3	3	3	–
A ^c	Pa·n s ⁻¹	8.57 × 10 ⁻²⁸	8.57 × 10 ⁻²⁸	8.57 × 10 ⁻²⁸	8.57 × 10 ⁻²⁸	8.57 × 10 ⁻²⁸	5.78 × 10 ⁻²⁷	5.78 × 10 ⁻²⁷	1.76 × 10 ⁻¹⁴	1.76 × 10 ⁻¹⁴	1.76 × 10 ⁻¹⁴	1.76 × 10 ⁻¹⁴	1.76 × 10 ⁻¹⁴	–
Q	kJ mol ⁻¹	223	223	223	223	223	485	485	430	430	430	430	430	–
V [*]	m ³ mol ⁻¹	0	0	0	0	0	0	0	11 × 10 ⁻⁶	11 × 10 ⁻⁶	11 × 10 ⁻⁶	11 × 10 ⁻⁶	11 × 10 ⁻⁶	–
Thermal parameters														
Specific heat	m ² s ⁻² K ⁻¹	750	750	750	750	750	750	750	1250	1250	1250	1250	1250	1250
Thermal conductivity	W m ⁻¹ K ⁻¹	2.25	2.25	2.25	2.25	2.25	2.25	2.25	5–2.4 ^e	5–2.4 ^e	5–2.4 ^e	5–2.4 ^e	52.5 ^f	52.5 ^f
Thermal diffusivity	m ² s ⁻¹	1 × 10 ⁻⁶	1 × 10 ⁻⁶	1 × 10 ⁻⁶	1 × 10 ⁻⁶	1 × 10 ⁻⁶	1 × 10 ⁻⁶	1 × 10 ⁻⁶	1.3–0.64 × 10 ^{-6 e}	1.3–0.64 × 10 ^{-6 e}	1.3–0.64 × 10 ^{-6 e}	1.3–0.64 × 10 ^{-6 e}	1.4 × 10 ⁻⁵	1.4 × 10 ⁻⁵
Thermal expansion coefficient	K ⁻¹	3 × 10 ⁻⁵	3 × 10 ⁻⁵	3 × 10 ⁻⁵	3 × 10 ⁻⁵	3 × 10 ⁻⁵	3 × 10 ⁻⁵	3 × 10 ⁻⁵	3.2–3.9 × 10 ^{-5g}	3.2–3.9 × 10 ^{-5g}	3.2–3.9 × 10 ^{-5g}	3.2–3.9 × 10 ^{-5g}	3.2–3.9 × 10 ^{-5g}	3.2–3.9 × 10 ^{-5g}
Radioactive heat production	μW m ⁻³	1	2	2	2	2	0.4	0	0	0	0	0	0	0

^a High-pressure (HP) and ultrahigh-pressure (UHP) densities.

^b Viscous scaling factor for accreted terrane lithospheric mantle varies among sensitivity experiments.

^c Experimental pre-exponential factor has been converted from the laboratory value to the corresponding one for the flow law written in terms of the second invariants of stress and strain. This is required so that the flow law remains the same in any reference frame.

^e Thermal conductivity and thermal diffusivity for mantle lithospheres vary linearly over the temperature range 273–1050 K.

^f Enhanced conductivity and thermal diffusivity required to maintain adiabatic temperature gradient of 0.4 K km⁻¹.

^g Thermal expansion coefficient increases linearly over the temperature range 500–2000 K.

^h Includes the effects of strain softening (see text and Appendix A).