The Dynamics of the India-Eurasia Collision: Faulted Viscous Continuum Models Constrained by High-Resolution Sentinel-1 InSAR and GNSS Velocities

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Key Points:

\begin{itemize}
  \item A suite of faulted viscous shell models testing key parameters explain new observations from geodesy for the India-Eurasia collision.
  \item The India-Eurasia collision is explained by the balance between buoyancy and boundary forces, slip-resistance on major faults, and internal viscosity variations.
  \item Central Tibetan Plateau has a vertically-averaged effective viscosity of $\sim 10^{21}$ Pa s, 1-2 orders lower than the surrounding area.
\end{itemize}

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Abstract

The dynamics of lithospheric deformation in the India-Eurasia collision zone has been debated over many decades. Here we test a two-dimensional (2-D) Thin Viscous Shell (TVS) approach that has been adapted to explicitly account for displacement on major faults and investigate the impact of lateral variations in depth-averaged lithospheric strength. We present a suite of dynamic models to explain the key features from new high-resolution Sentinel-1 Interferometric Synthetic Aperture Radar (InSAR) as well as Global Navigation Satellite System (GNSS) velocities. Comparisons between calculated and geodetically observed velocity and strain rate fields indicate: (a) internal buoyancy forces from Gravitational Potential Energy (GPE) acting on a relatively weak region of highest topography (>2,000 m) contribute to dilatation of the high plateau and contraction on the margins; (b) a weak central Tibetan Plateau (~10^21 Pa s compared to far-field depth-averaged effective viscosity of 10^{22} to 10^{23} Pa s) is required to explain the observed long-wavelength eastward velocity variation away from major faults; (c) slip on faults produces strain localization and clockwise rotation around the Eastern Himalayan Syntaxis (EHS). We discuss the tectonic implications for rheology of the lithosphere, distribution of geodetic strain, and partitioning of active faulting and seismicity.

Plain Language Summary

The collision of the Indian Plate with Eurasia has created the Tibetan Plateau, the largest deforming region in the continents. It has been a focus for heated debate and has inspired two contrasting tectonic models: (a) The deformation is localized on major faults separating “blocks” or (b) the strain is distributed throughout a “continuum”. We approximate the India-Eurasia collision by treating the continent as a two-dimensional viscous fluid with regional variations in strength, explicitly accounting for displacements on selected major faults. We present a suite of models to explain the key features of new observations from satellites. The best-fit model involves a weak Tibetan Plateau, a particularly weaker central Plateau, and four strong regions outside the Plateau, and requires resistance to slip on faults. This represents the deformation field of the India-Eurasia collision zone as a combination of continuous distributed deformation and focused strain on major faults.

1 Introduction

The Tibetan Plateau was created by the collision of the Indian Plate with Eurasia and has long been a testing ground for models of continental deformation. It extends more than 2,000 km north of the Himalayan Frontal Thrust, where large active faults appear to have developed since middle Miocene (Duvall et al., 2013). Geodetic observations from Global Navigation Satellite System (GNSS) and Interferometric Synthetic Aperture Radar (InSAR) reveal a complex pattern of current deformation in the India-Eurasia collision zone (Figure 1). The Tibetan Plateau and its margins accommodate India’s indentation into Eurasia by crustal shortening, widespread active faulting, folding and uplifting (Q. Wang et al., 2001). In the Eurasia fixed reference frame, the westward motion in the western Tibetan Plateau is tapered to zero while the eastward velocities increase over ~1,000 km distance across the eastern plateau before decreasing rapidly outside the plateau (M. Wang & Shen, 2020). Deformation within the plateau and the Tian Shan to the north is broadly distributed whereas outside these areas there are large undeforming regions with deformation focused around the perimeter of these regions (Ge et al., 2015; Zheng et al., 2017; M. Wang & Shen, 2020; W. Li et al., 2022). One of these undeforming regions, the Tarim Basin between the plateau and Tian Shan, has been observed to rotate clockwise at a rate of 0.4-0.6°/Myr with respect to Eurasia since the Cenozoic era (Avouac & Tapponnier, 1993; Z.-K. Shen et al., 2001; Craig et al., 2012; J. Zhao et al., 2019; M. Wang & Shen, 2020).
Figure 1. (a) Eastward velocity map constructed from ascending and descending Sentinel-1 InSAR line-of-sight velocities (Wright et al., 2023). Black lines show the location of profiles presented in Figure 6. Dark red lines are fault traces from the Global Earthquake Model (GEM) Global Active Faults Database (Styron & Pagani, 2020). Thick black lines are model faults incorporated in numerical simulations in this study. Abbreviation of names for fault profiles: KA = Karakoram Fault, KK = Karakash Fault, LGC = Longmu-Gozha Co Fault, ATF = Altyn Tagh Fault, HY = Haiyuan Fault, KL = Kunlun Fault, XSH = Xianshuihe Fault. (b) GNSS velocities (Wright et al., 2023, and references therein). Purple polygon shows the boundary of calculation domain. Dashed lines show the extents of zoomed view shown in Figure 12. Thick black lines are model faults incorporated in this study.
While some major strike-slip faults in the Tibetan Plateau show strain concentrations (Kreemer et al., 2014; Ge et al., 2015), there are also areas of diffuse strain (Zheng et al., 2017). The high plateau is dilating at a rate of $\sim 10^{-20}$ nanostrain/yr (Molnar & Deng, 1984; Ge et al., 2015; Zheng et al., 2017; Wright et al., 2021, 2022, 2023). WNW-ESE extension occurs throughout the plateau interior through a set of north-south striking rifts/grabens and conjugate strike-slip faulting (Molnar & Tapponnier, 1978; Duvall et al., 2013; H. Wang et al., 2019); the northern and southern regions of the Tibetan Plateau show similar rates of dilatation in short-term geodetic data (Ge et al., 2015), although geological data suggest arc-parallel extension rates in the plateau may be higher nearer the Himalayan arc (Copley et al., 2011). The northeastern Tibetan Plateau and the eastern and southern margins of the plateau, as well as the Tian Shan region, are experiencing rapid contraction (Molnar & Tapponnier, 1978; Q. Wang et al., 2001; England & Molnar, 2015; Y. Li et al., 2018; Metzger et al., 2020, 2021; J. Li et al., 2022; Ou et al., 2022; Zhu et al., 2022). The southeastern Tibetan Plateau rotates clockwise around the eastern Himalayan syntaxis (EHS) (Q. Wang et al., 2001; Z. Shen et al., 2005; W. Wang et al., 2017; Zheng et al., 2017; Y. Li et al., 2019; M. Wang & Shen, 2020; Gan et al., 2021; W. Wang et al., 2021).

How best to understand the deformation field produced by the India-Asia collision has been a subject of extensive debate (Thatcher, 2009; Searle et al., 2011; Bendick & Flesch, 2013; P. Zhang, 2013; H. Zhang et al., 2020; Dal Zilio et al., 2021). Since the early days of plate tectonics, which beautifully explains the motion of oceanic plates, it has been recognized that deformation of the continents cannot be described by the motion of only a few large plates, with seismicity focused around their edges (McKenzie, 1972). Nevertheless, a popular approach for characterizing continental deformation is to model the deformation as rotation and translation of a number of blocks, or microplates, each following the kinematic rules of plate tectonics (Avouac & Tapponnier, 1993; McCaffrey et al., 2000; McClusky et al., 2001; Wallace et al., 2004; Meade & Hager, 2005; Wallace et al., 2005; Socquet et al., 2006; Thatcher, 2007; W. Wang et al., 2017; Y. Li et al., 2018; W. Wang et al., 2021). In most formulations of block models, strain concentrations only occur along the block boundaries, although a few allow for strain within block interiors (Q. Chen et al., 2004; Loveless & Meade, 2011). Avouac and Tapponnier (1993) proposed the first 4-microplate model for the India-Asia collision based primarily on geological observations. Q. Chen et al. (2004) constructed a deformable block model to explain GNSS observations from 45 stations. The trend in subsequent models has been to increase the number of blocks to fit more GNSS observations as they become available (Thatcher, 2007; Loveless & Meade, 2011; W. Wang et al., 2017; Y. Li et al., 2018; W. Wang et al., 2021; Styron, 2022). These block models are helpful for deriving slip rates and locking depths for major faults and are widely used in seismic hazard analysis (W. Wang et al., 2017; Y. Li et al., 2018; W. Wang et al., 2021; Styron, 2022). They can naturally describe large undeforming areas and focused strain around faults. If enough blocks are used, these models can reproduce any observed features of the strain field. However, because the models are purely kinematic, they have no predictive power and cannot be used to test the underlying causes of the observed deformation or to understand the balance of forces acting on blocks. The geodetic strain can be described in the short term, even with an elastic model, but appealing to elastic strain as an explanation of strain rates sustained on geological time-scales is not logically self-consistent. In addition, as focused strain might not coincide with mapped faults (H. Wang & Wright, 2012; H. Wang et al., 2019), a simple block model could underestimate the likelihood of earthquakes occurring on unknown faults due to our imperfect knowledge of the boundaries of crustal blocks, which must be defined a priori; all earthquakes by definition must occur on block boundaries in such a framework.

An alternative approach has been to treat continents as a continuum, with deformation modeled as a viscous fluid acting under the influence of the internal and boundary forces applied, and a simply parameterized viscous constitutive law (England & McKenzie, 1997).
zie, 1982; Flesch et al., 2001). In these models, deformation is distributed throughout
the layer representing the lithosphere. England and McKenzie (1982) simplified the de-
formation to a two-dimensional (2-D) problem by treating the lithosphere as a thin vis-
cose sheet originally developed for a flat layer with vertically-averaged properties. England
and Houseman (1986) applied the viscous sheet formulation assuming a uniform viscous-
ity coefficient to analyze the dynamics of the India-Eurasia collision. In such models, strain
is focused where gradients of Gravitational Potential Energy (GPE) are greatest, and
on parts of the boundary where the boundary forces change rapidly. With more and more
observations and stronger computational power, more complexity in models has been re-
quired to explain the observations (Neil & Houseman, 1997; Flesch et al., 2001; Vergnolle
et al., 2007; Lechmann et al., 2014; Bischoff & Flesch, 2018, 2019). Early viscous con-
tinuum models did not predict the strain concentrations observed in dense geodetic data
around major faults. However, Dayem et al. (2009) and Molnar and Dayem (2010) showed
that viscous continuum models can concentrate strain at regions of strength contrast.
Lechmann et al. (2014) and Bischoff and Flesch (2018, 2019) achieved strain concentra-
tions by explicitly allowing weaker regions to represent localized strain associated with
major faults.

The lower crust is expected to be relatively weak based on typical power law creep
laws (Brace & Kohlstedt, 1980). Some authors have argued that the lower crust is so weak
that it is decoupled from both the upper crust and the upper mantle. W. Zhao and Mor-
gan (1987) presented a model in which the stronger Indian crust injects into the weaker
fluid-like lower crust of the Tibetan Plateau. Based on geologic and GNSS observations,
Royden et al. (1997, 2008) presented a lower crustal flow model in the eastern Tibetan
Plateau where crustal material flows around the EHS and also around the strong Sichuan
Basin. They argued that the lower crust escapes from beneath the central plateau through
regions where crust is weak (Clark & Royden, 2000), and that the morphology of the east-
ern plateau reflects crustal material flows. Copley and McKenzie (2007) interpreted the
formation of the geometry of the EHS by gravitationally driven fluid flow in both the
southern Tibetan Plateau and the Indo-Burman Ranges. Bischoff and Flesch (2019) ap-
proximated the three-dimensional (3-D) India-Eurasia deformation with creeping flow,
with a weak lower crust required to explain the observed vertical surface velocities. How-
ever, Rey et al. (2010) show that large-scale relative displacement of the lower and up-
per crust is unlikely. Their result justifies a key assumption of the TVS method that the
lithosphere deforms coherently with depth, that is, horizontal velocity is independent of
depth and horizontal tractions can be vertically averaged.

Lower crustal channel flow has also been invoked for models in which material in
a partially molten mid-crust is extruded southward from beneath the southern Tibetan
Plateau towards the high Himalayan slab (Grujic et al., 2002; Searle et al., 2003; Law
et al., 2004; Searle & Szulc, 2005; Godin et al., 2006; Searle et al., 2006, 2011). Beaumont
et al. (2001) interpreted the Himalayan tectonics by a low-viscosity channel flow and duc-
tile extrusion; high-grade metamorphic rocks were exhumed from this channel. However,
Copley et al. (2011) argued that the mechanical coupling between the upper crust of the
southern Tibetan Plateau and the underthrust Indian crust is inconsistent with the low-
viscosity ‘channel flow’ models in the southern plateau. Flesch et al. (2018) suggest sur-
face GNSS velocities contain little or no information about 3-D dynamics. Penney and
Copley (2021) further suggest that the temporal evolution of topography in the south-
eastern Tibetan Plateau can be explained without invoking a low-viscosity lower crustal
channel.

Both block models and continuum models are over-simplifications of a more com-
plex reality that requires both distributed deformation and, at least in the near surface,
slip on faults (Thatcher, 2009). Ductile deformation is manifested in almost any geolog-
ical environment where the temperatures are sufficiently great, but near surface defor-
mation typically occurs by faulting. In the case of large-scale continental faults, seismic
activity is typically restricted to the upper 15 km or so (Wright et al., 2013), but there is increasing evidence that localized deformation is moderated by ductile shear zones that can extend through the crustal layer and possibly into the mantle (Warner, 1990; Kellen- men & Dick, 1995; Leloup et al., 1999; Thybo et al., 2000; Bürgmann & Dresen, 2008; Vauche et al., 2012; Alivizuri & Hetényi, 2019; Scholz & Choi, 2022). Hence the deformation field in general can be represented as a continuum modulated by major faults. Continuum models are appealing in that they have the potential to explain large-scale deformation with relatively few adjustable parameters. Garthwaite and Houseman (2011) demonstrate the validity of the 2-D thin viscous sheet approximation for continental collision provided that the indenter width is larger than the thickness of the lithosphere. In this study, we employ the adapted 2-D TVS continuum model of England and McKenzie (1982), explicitly modified to account for displacement discontinuities on faults. Although a linear constitutive relation between stress and strain rate is often adopted in 3-D numerical modeling (Royden et al., 1997; F. Shen et al., 2001; Liu & Yang, 2003; Copley & McKenzie, 2007; Lechmann et al., 2014; Bischoff & Flesch, 2019; Penney & Copley, 2021), we assume a non-Newtonian (power law) viscous rheology. Early geodynamic simulations have primarily relied on information from topography, Quaternary fault slip rates, and seismic moment tensors. The constantly-improving accuracy and resolution of the geodetic observations now enable tighter constraints on tectonic models. We present a suite of faulted viscous continuum models constrained by new geodetic observations of the India-Eurasia collision. This allows us to explore (a) the importance of internal buoyancy forces from GPE, (b) the relationship between slip resistance on faults and associated ductile deformation, and (c) the role of rheological/strength contrasts and how they modulate and localize deformation.

2 Data and Methods

2.1 Data

We use constraints from new high-resolution InSAR and published GNSS horizontal velocity fields (Wright et al., 2023, and references therein, Figure 1) to test the faulted viscous continuum model. Both datasets are fixed to a Eurasia reference frame. As relative motion across the Himalaya and Indo-Burma subduction zones appears to be controlled by 3-D geometry (Ni et al., 1989; C. Li et al., 2008b; Liang et al., 2016; Dubey et al., 2022), we do not incorporate measurements within the Indian Plate in our 2-D dynamic modeling. We obtain a relatively sparse set of velocity vectors by a weighted average of joint model velocities of InSAR and GNSS (Wright et al., 2023) derived from the VELMAP approach (H. Wang & Wright, 2012). We sub-sample the combined geodetic solution onto a 1° (longitude) by 0.5° (latitude) grid using a Gaussian weight of all samples within 0.5° distance. The half-width at half-height of the Gaussian weight function is 0.593°. We produce a total of 232 points for the combined geodetic observations at 2° × 1° spacing in longitude and latitude (blue arrows in Figure 2c). We also test our models using a more extensive set of horizontal GNSS measurements (Wright et al., 2023, and references therein, Figure 1b). Excluding GNSS measurement points that are too close to model faults (<10 km) or too close together (<10 km) and have greater uncertainty, we use 2,656 GNSS measurements as constraints.

2.2 Methods

2.2.1 Power Law Rheology in a Faulted Ductile Medium

The vertically-averaged rheology of the TVS is described by a power law relation between deviatoric stress and strain rate (England & McKenzie, 1982; Sonder & England, 1986):

\[ \tau_{ij} = B \dot{E}^{(\frac{1}{n}-1)} \dot{\varepsilon}_{ij} \]  

(1)
where $\tau_{ij}$ is the $ij$th component of the deviatoric stress (averaged over the thickness of the lithosphere, $L$), $\dot{\varepsilon}_{ij}$ is the $ij$th component of the strain rate tensor (assumed constant with depth), and $\dot{E}$ is the second invariant of the strain rate tensor:

$$\dot{E} = \sqrt{\dot{\varepsilon}_{kk} \dot{\varepsilon}_{kk}}$$

(2)

The fluid is assumed to be incompressible ($\varepsilon_{kk}=0$). The viscosity coefficient, $B$, and the power law exponent, $n$, define the physical properties of the lithosphere. In this study, we use $n=3$, which is suitable for a lithosphere where depth-averaged rheology is dominated by the power law creep of olivine (Brace & Kohlstedt, 1980; Karato et al., 1986; Kirby & Kronenberg, 1987), whereas large $n$ represents plastic behavior (Goetze et al., 1978). The effective viscosity is

$$\eta_{\text{eff}} = \frac{1}{2} B \dot{E}^{(\frac{1}{2} - 1)}$$

(3)

Note that for non-Newtonian fluids ($n \neq 1$) the effective viscosity is dependent on strain rate. The GPE is calculated assuming local isostatic balance of topography ETOPO1 (Amante & Eakins, 2009) smoothed with a Gaussian filter width of 20 km. The Argand number, $Ar$, as defined by England and McKenzie (1982), represents the relative importance of gravitational buoyancy related stress to viscous stress required to deform the lithosphere at a reference strain rate $U_0 L$:

$$Ar = \frac{g \rho_c L (1 - \frac{\rho_m}{\rho_c})}{B_0 (\frac{L}{R})^{\frac{1}{n}}}$$

(4)

where $g$ is the gravitational acceleration, $\rho_c$ and $\rho_m$ are the average densities of crust and mantle, respectively, $B_0$ is the scale factor for the viscosity coefficient, and $U_0$ is a scale velocity determined by minimizing the root mean square (RMS) misfit function:

$$M = \left[ \frac{1}{N} \sum_{i=1}^{N} |u_i - U_0 u_i'|^2 \right]^{\frac{1}{2}}$$

(5)

where $u_i$ is the $i$th observed velocity, and $u_i'$ is the dimensionless velocity of the same site in the calculation. In the dimensionless force balance, the Argand number multiplies the lateral gradient of GPE, scaling the force that pushes the layer away from regions of high GPE.

We assume that the continuum deformation may be interrupted by slip on model fault structures, with resistance to displacement proportional to the slip rate for tractions and displacements in the horizontal plane. The depth-averaged shear traction for these model faults is assumed dominated by the behavior of ductile shear zones beneath the seismically active layer. Therefore, we assume for tangential ($\sigma_t$) and normal ($\sigma_n$) directions:

$$\sigma_t = f'_t \Delta U$$

(6)

$$\sigma_n = f'_n \Delta U$$

(7)

where $f'_t$ and $f'_n$ represent the dimensionless fault-resistance coefficients in tangential and normal directions, respectively, with zero implying a free-slipping fault and infinity meaning a locked fault. The fault-resistance coefficient has dimensions of stress/velocity, depending on the choice of $Ar$. Its scale factor is

$$f_0 = \frac{B_0 (\frac{L^2}{R})^{\frac{1}{2}}}{U_0}$$

(8)

where $R$ is the radius of the Earth.

We explicitly allow for displacement discontinuities across major faults (Altyn Tagh, Haiyuan, Kunlun, Xianshuihe, Sagaing, Main Pamir Thrust faults, and eastern boundary of the Indian Plate) in the India-Asia collision zone where InSAR and GNSS reveal apparent velocity contrasts (Figure 1).
2.2.2 Boundary Conditions and Internal Structures

We use the adapted finite element code BASIL (Houseman et al., 2002) for numerical modeling. The program solves the stress-balance equations using the finite-element method described by Houseman and England (1986) amended to represent a deformation field on a spherical shell, as used by England et al. (2016). Figure 2a shows the boundary conditions. We set velocities to zero along the northern, western, and part of southern boundaries which are assumed fixed to the undeforming Eurasian plate ($U_E = U_N = 0$). We set plate rotations on three boundary sections; we use the reconstructed motion of the Indian Plate relative to Eurasia (IND-EUR) from DeMets et al. (2020) and MORVEL velocities of Yangtze (YZ-EUR) and Amur Plates (AM-EUR) from DeMets et al. (2010). We set the rotation rate of the Indian Plate to 1 (dimensionless) and scale those of Yangtze and Amur Plate boundary segments in proportion. The velocity scale $U_0$ is determined from the solution by minimizing the misfit (Eq. 5) between observed and dimensionless model velocities. The velocities on the part of the southern boundary that crosses Myanmar are poorly constrained and we set zero velocity in the east direction and zero traction (relative to lithostatic) in the north direction ($U_E = T_N = 0$, Figure 2a); this allows for normal motion along that segment as implied by GNSS measurements in that region (Figure 1b). The complexity of the observed deformation styles indicates the convergence of India with Eurasia is not the only factor influencing the distribution of displacements. The internal buoyancy forces from GPE and heterogeneities in lithospheric strength also contribute to the regional deformation pattern (England & Houseman, 1985; England & Molnar, 2005). Assuming that the background dimensionless depth-averaged viscosity coefficient ($B'$) is 1, we also investigate the influence of regional variations in internal strength by embedding strong Indian Plate, Tarim, Sichuan, and Alxa-Ordos Basins ($B'_S=10$) (Figures 2a and 3a), weakening ($B'_W < 1$) area of high topography defined by the contour of $\sim 2,000$ m elevation and bounding faults (Figure 4a), and/or central Tibetan Plateau (Figure 5a).

3 Numerical Simulations and Results

We conduct a comprehensive suite of numerical experiments, aiming to match the key features of the geodetic observations (Table 1) under a fixed set of boundary conditions. We incrementally build up the complexity of models in terms of the number of features employed, with the aim to find the most parsimonious solution that matches the large-scale, systematic patterns of the velocity field. In Case 1, we investigate internal strength variations by involving strong Indian Plate, Tarim, Sichuan, and Alxa-Ordos Basins, a weak area of high topography, and/or a weak central Tibetan Plateau. In Case 2, we account for displacement discontinuities by explicitly incorporating faults. In each case we explore the parameter space systematically to obtain a minimum RMS misfit between observations and model horizontal velocities. We compare observed and model gridded eastward velocities for all the experiments as InSAR observations are almost insensitive to north-south motion. However, the InSAR velocity field is also constrained by GNSS measurements of the north component of velocity and our measures of model misfit are equally weighted in both components.

3.1 Case 1: Lateral Heterogeneity in Viscosity Coefficient

3.1.1 Case 1.1: Rigid India Indenter

In this case, we simulate the convergence of India with Eurasia by embedding a rigid Indian Plate in the otherwise homogeneous model domain (Figure 2a). Doing so allows us to apply the present rotation rate vector for India relative to Eurasia (Section 2.2.2) to the arbitrary southern boundary of the domain, in order to produce the apparent motion of the relatively rigid Indian Plate. The depth-averaged viscosity coefficient is set to 10. Because $n = 3$, setting $B'_S=10$ can result in strain rates $10^3$ times smaller than
Table 1. Summary of Model Cases to Match the Key Observable Features of the Geodetically-Derived Velocity Field in the India-Asia Convergence Zone

<table>
<thead>
<tr>
<th>Key observations</th>
<th>Case 1: Lateral heterogeneity in viscosity coefficient</th>
<th>Case 2: Allowing displacements on selected major faults</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Case 1.1: Rigid India indenter</td>
<td>Case 1.2: Embedding strong Indian plate, Tarim, Sichuan, and Alxa-Ordos basins</td>
</tr>
<tr>
<td></td>
<td>Case 1.3: Weakening area of high topography</td>
<td>Case 1.4: Weakening central Tibetan Plateau</td>
</tr>
<tr>
<td>Distributed deformation throughout the India-Eurasia collision zone</td>
<td>✓</td>
<td>✓</td>
</tr>
<tr>
<td>Dilatation of high plateau</td>
<td>×</td>
<td>✓</td>
</tr>
<tr>
<td>Contraction on the margins of plateau</td>
<td>Partly</td>
<td>Partly</td>
</tr>
<tr>
<td>Smooth, long-wavelength eastward velocity variation away from major faults</td>
<td>×</td>
<td>×</td>
</tr>
<tr>
<td>Strain concentrations on major faults</td>
<td>×</td>
<td>×</td>
</tr>
<tr>
<td>Asymmetric eastward velocity gradient across the Tibetan Plateau</td>
<td>×</td>
<td>×</td>
</tr>
<tr>
<td>Clockwise rotation around the EHS</td>
<td>×</td>
<td>×</td>
</tr>
<tr>
<td>Clockwise rotation of the Tarim basin (rotation rate, °/Myr)</td>
<td>-0.149</td>
<td>-0.161</td>
</tr>
<tr>
<td>Best-fit Argand number</td>
<td>1.0</td>
<td>1.8</td>
</tr>
<tr>
<td>RMS misfit</td>
<td>6.7</td>
<td>6.6</td>
</tr>
</tbody>
</table>

*a*The rotation rate is calculated based on model GNSS velocities within the Tarim block for each case, anti-clockwise positive. The rotation rate of the Tarim basin derived from GNSS observations (Figure 1b) is -0.592 °/Myr.

*b*RMS misfit to joint model horizontal velocities of InSAR and GNSS (mm/yr)
in an adjoining region where $B' = 1$, though the effect of irregular geometry makes for a more complex dependence of strain-rate on $B'_S$. The Argand number $A_r = 1$ gives the minimum RMS misfit (6.7 mm/yr, Figure 2b) subject to the choice of $n = 3$ and specified boundary conditions. No displacement is allowed on faults but we observe strain concentrations on the syntaxial regions on either end of the Himalayan chain, and also at points on the external boundary of the domain (Figure S1a), where there is an abrupt change in the boundary conditions. This calculation produces subtle E-W extension/dilatation in the Tibetan Plateau where the ratio of E-W extension rate to N-S convergence rate is around 0.1 (Figures 2c, S1a, and S1b). Clockwise rotation around the EHS is not reproduced (Figure 2c).

3.1.2 Case 1.2: Embedding Strong Indian Plate, Tarim, Sichuan, and Alxa-Ordos Basins

Based on the coherent displacement patterns of the Indian Plate, Tarim, Sichuan, and Alxa-Ordos Basins observed in GNSS dataset, these lithospheric blocks are interpreted to behave as rigid blocks with relatively cold thermal profiles (Tapponnier & Molnar, 1976; Kao et al., 2001; Q. Wang et al., 2001; Yang & Liu, 2002; Jagadeesh & Rai, 2008; C. Li et al., 2008a; P. Zhang & Gan, 2008; Z. Zhang et al., 2010; Craig et al., 2012; Mahesh et al., 2012; C.-L. Zhang et al., 2013; Deng & Tesauro, 2016; Rui & Stamps, 2016). We investigate the impact of involving the four rheologically strong regions, with a viscosity coefficient one order of magnitude higher than background ($B'_S = 10$) (Figure 3a). The outlines of the rigid regions are approximated from the surface geomorphology/topography.
Figure 3. Same as Figure 2, but for Case 1.2: embedding strong Indian Plate, Tarim, Sichuan, and Alxa-Ordos Basins.

In this calculation, the minimum misfit (∼6.6 mm/yr) obtained for Argand number 1.8 (Figure 3b) is comparable to Case 1.1. Asymmetric eastward velocity gradient in the western and eastern Tibetan Plateau and clockwise rotation around the EHS are not recovered (Figure 3c). Negligible strain occurs in the interiors of the rigid blocks (Figures S1c and S1d) and northward displacement rate vectors are still predominant everywhere in the solution domain in contrast to observed eastward rates in the eastern Tibetan Plateau.

3.1.3 Case 1.3: Weakening Area of High Topography

The lithosphere of the Tibetan Plateau and Tian Shan has been suggested to be relatively thinner, hotter and rheologically weaker than the indenting Indian Plate (Tapponnier & Molnar, 1979; Molnar & Tapponnier, 1981). In this case we explore the effect of such weakening regions of high elevation. We choose the shape of the weak region to follow approximately the smoothed contour of ∼2,000 m topography bounded by faults in places (medium blue zone in Figure 4a). We search for an optimal combination of the Argand number and the viscosity coefficient of the weak zone ($B'_W$). A minimum misfit of 5.9 mm/yr was obtained with $Ar$ of ∼3.5 and $B'_W$ of ∼0.4 (Figure 4b), indicating that gravitational spreading plays a more significant role when enabled by weakened thick crust. It can be seen that there is some trade-off between $Ar$ and $B'_W$: as $Ar$ increases, a relatively ‘stronger’ weak zone would be required. This model calculation enhances the expression of eastward motion in the eastern Tibetan Plateau (Figures 4c and S2c). Clockwise rotation around the EHS is still missing (Figure 4c). Note that strain becomes concentrated at regions of strength contrast; this experiment yields nearly E-W extension throughout much of the central-southern Tibetan Plateau and NNW-SSE stretching around the EHS (Figure S1e). The high plateau is dilating, as the weaker plateau is enabled to flow outward from the region of high GPE. The margins of the plateau show convergence (Figure S1f). These patterns are broadly consistent with the geodetically-derived dilatation strain rate field (Wright et al., 2023).
Figure 4. (a) Schematic diagram illustrating boundary conditions and model rheological coefficients for Case 1.3: weakening area of high topography. The weak zone follows the contour of \( \sim 2,000 \) m elevation bounded by faults in places. (b) Misfit as a function of the Argand number and viscosity coefficient of the weak zone. The minimum misfit is marked as star. Conventions of (c) and (d) are as described in Figure 2.
3.1.4 Case 1.4: Weakening Central Tibetan Plateau

We note that none of the above experiments can produce the observed long-wavelength increase in eastward velocity across the Tibetan Plateau (Figure 6a). We now include in the model an additional rheologically weak central plateau roughly following the shape of the commonly referenced Qiangtang Block (Liu & Yang, 2003; P. Zhang et al., 2003), which is bounded by the Jinsha Suture-Kunlun Fault-Xianshuihe Fault to the north, the Bangong-Nujiang Suture-Jiali Fault-Red River Fault to the south, part of the Karakoram Fault to the west, and the northwestern boundary of the Dianzhong Block to the east (dark blue zone in Figure 5a). As the Dianzhong Block appears to obstruct the material extrusion to the southeast (Han et al., 2022), we exclude the Dianzhong Block from the weak region and keep its viscosity coefficient as that of the background. The misfit is dependent on the Argand number, viscosity coefficients of the weak high topographic area ($B'_{W1}$) and central Tibetan Plateau ($B'_{W2}$). The combination of the three parameters (4.0, 0.5, 0.1, respectively) leads to a minimum misfit of 4.9 mm/yr (Figure 5b), as opposed to 5.9 mm/yr in Case 1.3. This simulation facilitates the eastward velocity gradient across the Tibetan Plateau (Figures 5c, 6a, and S2d). Again, the clockwise rotation around the EHS is not reproduced (Figure 5c). The strain rate fields in this calculation are similar to those of Case 1.3, except for additional strain concentration at regions of strength contrasts (Figures S1g and S1h). The significance of this experiment is that we recover the gradient of eastward velocity across the Tibetan Plateau ($\sim$20 mm/yr contrast over $\sim$1,400 km distance, compared to $\sim$10 mm/yr difference over that distance in Case 1.3, Figure 6a).

3.2 Case 2: Allowing Displacements on Selected Major Faults

In Cases 1.3 and 1.4, strain is concentrated at regions of strength contrast (Figures S1e and S1g). As obvious velocity gradients have been observed across major faults in the Tibetan Plateau (Wright et al., 2023, Figure 6), we introduce strain localization on faults by explicitly allowing for displacement discontinuities across the faults in Case 2.

3.2.1 Case 2.1: Absence of Weak Zone

We first exclude any weak regions to investigate the impact of fault-resistance coefficients. We take into account the dominant strike-slip motion along major faults (Altyn Tagh, Haiyuan, Kunlun, and Xianshuihe Faults) by applying a constant strike-parallel fault-resistance coefficient ($f'_t$) along a fault. We also allow dip-slip motion on the eastern boundary of the Indian Plate, the Sagaing Fault, and the Main Pamir Thrust Fault by applying $f'_n$ parameters simultaneously. Model faults are delineated as thick black lines in Figure 7a. In this case, we allow $Ar$ and $f'_t$ to be free parameters. To maintain the simplicity of the calculations, we assume a uniform $f'_t$ for all model faults (with the same-magnitude $f_n$ applied for the Sagaing Fault), and set $f'_n = 10$ for both the eastern boundary of the Indian Plate and the Main Pamir Thrust Fault determined by trial and error (Figure 7a). We obtained a minimum misfit of 5.1 mm/yr with $Ar$ of 7.4 (Figure 7b). The fault-resistance coefficients are at most $\sim$0.2 (Figure 7b), indicating that the faults tend to be free-slipping. This calculation allows discontinuities in the velocity component across faults (Figures 6 and S2e) and reproduces the asymmetric eastward velocity gradient across the Tibetan Plateau (Figure 7c). Relative to previous simulations, Case 2.1 predicts a greater rate of clockwise rotation of the Tarim Basin, owing to shear motion allowed on the Altyn Tagh Fault as its southern boundary. Clockwise rotation around the EHS is also enhanced, due to the model allowing local convergence on the Sagaing Fault and the eastern boundary of the Indian Plate as a rough representation of Indo-Burma subduction. The fault-resistance coefficients determine the velocity steps across the faults (Figure 6). Note that geodetic data constrain short-term interseismic strain rates across locked faults, whereas the geodynamic model is predict-
Figure 5. (a) Schematic diagram illustrating boundary conditions and model rheological coefficients for Case 1.4: weakening central Tibetan Plateau. QB = Qiangtang Block, JS = Jinsha Suture, KL = Kunlun Fault, XSH = Xianshuihe Fault, KA = Karakoram Fault, BNS = Bangong-Nujiang Suture, JL = Jiali Fault, RR = Red River Fault. The Dianzhong Block (DB) is delimited by dashed polygon, with viscosity coefficient of 1 as background. (b) RMS misfit as a function of the Argand number, viscosity coefficients of high topographic area ($B'_W$) and central Tibetan Plateau ($B''_W$). Stars denote the best fits for each value of $B''_W$ tested, with the global minimum misfit occurring at $B''_W = 0.1$. Conventions of (c) and (d) are as described in Figure 2.
Figure 6. Eastward velocity profiles whose locations and labels are shown in Figure 1a. Velocities from InSAR (within 40 km bin) and GNSS (within 100 km bin) observations are shown as gray dots and cyan dots with 1-sigma error bars, respectively (Wright et al., 2023, and references therein). Yellow bars mark the location of faults. (a, b) Two long profiles nearly perpendicular or parallel to the direction of the India-Asia convergence. Colored lines represent model velocities for each case, among which cases without faults are shown as dashed lines while cases with faults are shown as solid lines. (c-n) Profiles across major strike-slip faults in the Tibetan Plateau showing the effect of the fault-resistance coefficients. Model velocities in cases without faults are shown as orange dashed lines ($f_t = \infty$). Pink lines denote faults that are free-slipping ($f_t = 0$). Dark red lines represent faults with uniform resistance to slip ($f_t = 0.4 \text{ MPa-yr/mm}$) for all model faults. Black lines show larger slip-resistance ($f_t = 3.5 \text{ MPa-yr/mm}$) for both ATF (including KK and LGC branches) and XSH, with $f_t = 0.4 \text{ MPa-yr/mm}$ for HY and KL. Faults: KK = Karakash Fault, LGC = Longmu-Gozha Co Fault, ATF = Altyn Tagh Fault, HY = Haiyuan Fault, KL = Kunlun Fault, XSH = Xianshuihe Fault, KA = Karakoram Fault.
Case 2.1: Incorporating Faults

(a) (b) (c) (d)

MPT: $f_t' = 0$ and $f_n' = 10$
ATF: $f_t' = 0$
HY: $f_t' = 0$
KL: $f_t' = 0$
XSH: $f_t' = 0$
EIND: $f_t' = 0$ and $f_n' = 10$
SG: $f_t' = 0$ and $f_n' = 0$

Figure 7. (a) Schematic diagram illustrating boundary conditions and internal structures for Case 2.1: incorporating faults without high-elevation weak zones. Thick black lines denote model faults: MPT = Main Pamir Thrust Fault, ATF = Altyn Tagh Fault, HY = Haiyuan Fault, KL = Kunlun Fault, XSH = Xianshuihe Fault, SG = Sagaing Fault, EIND = eastern boundary of the Indian Plate. $f_t'$ and $f_n'$ are fault-resistance coefficients in tangential and normal directions, respectively. (b) Misfit as a function of the Argand number and fault-resistance coefficient. The best-fit solution has $A_r = 7.4$, $f_t' = 0$ and $f_n' = \text{inf}$ for all model faults, except $f_n' = 0$ for SG, $f_n' = 10$ for MPT and EIND. (c) Model fits (red arrows) to the sampled observations (blue arrows) from joint model velocities of InSAR and GNSS. Model faults are shown in thick black lines. (d) Misfit vectors.

3.2.2 Case 2.2: Embedding Weak Region of High Topography

We now include (Figure 8a) the weak high-elevation areas along with the faults, as described in Section 3.1.3, in attempting to reproduce the dilatation of high plateau and convergence on the margins of the plateau, especially in the northeastern plateau (Case 1.3, Figure S1f, and Table 1). For a given viscosity coefficient of the weak zone ($B'_W$ of 0.2, 0.4, 0.6, 0.8, and 1), we explore an optimal combination of the Argand number and fault-resistance coefficients. The model favors a $B'_W$ of 0.4 for the weak zone, $A_r$ of $\sim 4$ and $f_t'$ of 0.2, with a misfit of 3.8 mm/yr (Figure 8b). This calculation predicts a gentler eastward velocity gradient ($<15$ mm/yr contrast over a distance of $\sim 1,400$ km) compared to the case without weak zones (Figure 6a).

374 ing long-term velocities and strains averaged over multiple earthquake cycles. To facilitate comparison, we apply a Gaussian filter of width 100 km to the model velocity field to simulate the effect of interseismic locking before calculating the strain rate fields (Figure S3). In this simulation, strain concentrations on major faults and dilatation of high plateau are reproduced (Figures S1i and S1j). However, the NE-SW and nearly E-W convergences on the northeastern and eastern margins of the plateau, respectively, are missing (Figure S1j). The long-wavelength eastward velocity variation away from major faults also is not well captured (Figure 6a).
Figure 8. (a) Schematic diagram illustrating boundary conditions and internal structures for Case 2.2: incorporating faults and weak region of high topography. The weak zone is bounded approximately by the ~2,000 m elevation contour and the major faults. Thick black lines represent model faults: MPT = Main Pamir Thrust Fault, ATF = Altyn Tagh Fault, HY = Haiyuan Fault, KL = Kunlun Fault, XSH = Xianshuihe Fault, SG = Sagaing Fault, EIND = eastern boundary of the Indian Plate. $f'_t$ and $f'_n$ are fault-resistance coefficients in tangential and normal directions, respectively. (b) Misfit as a function of the Argand number, fault-resistance coefficient, and viscosity coefficient of the weak region. Well-matched parameter combinations are shown as stars, with the global minimum misfit occurring at $B'_W = 0.4$. The best-fit solution has $Ar = 4.0$, $f'_t = 0.2$ and $f'_n = \infty$ for all model faults, except $f'_n = 0.2$ for SG, $f'_n = 10$ for MPT and EIND. Conventions of (c) and (d) are as described in Figure 7.

3.2.3 Case 2.3: Weakening Central Tibetan Plateau

In this case, we present a hybrid model incorporating both faults and laterally varying viscosity coefficients. Case 1.4 shows weakened central Tibetan Plateau with $B''_{W1}$ of 0.1, which produces the observed smooth, long-wavelength eastward velocity variation across the plateau (Figure 6a). We here search for a best-fit combination of the Argand number, fault-resistance coefficient, and viscosity coefficient of the weak high topographic region ($B'_W$), with $B''_{W2}$ fixed at 0.1 (Figures 9a and 9b). The misfit was reduced to 3.5 mm/yr (Figure 9b). In comparison with Case 2.2, the main improvement of Case 2.3 is that the long-wavelength eastward velocity variation has been well captured, with ~20 mm/yr gradient over ~1,400 km (Figure 6a). The model eastward velocity field (Figure 10) and model-derived strain rate fields (Figure 11) show agreement with the geodetic observations (Figures 1 and 11, Wright et al., 2023). This simulation explains all the key features of the India-Eurasia convergence evident in the geodetic ob-
Figure 9. Same as Figure 8, but for Case 2.3: incorporating faults and further weakened central Tibetan Plateau. The best-fit solution has $Ar = 4.0$, $f_t' = 0.5$ and $f_n' = \inf$ for all model faults, except $f_n' = 0.5$ for SG, $f_n' = 10$ for MPT and EIND.

Discussion

4.1 Slip Resistance on Faults Embedded in a Viscous Continuum

A “fault” in the context of the TVS model represents localized strain that is mediated in part by slip on a near-surface fault and by viscous strain of a narrow ductile shear zone at greater depths. Deformation can be generally represented as a continuum influenced by faults. Continuum deformation may comprise both elastic (e.g., earthquakes) and ductile (e.g., folds and shear zones) behavior. The elastic deformation may be ne-
Figure 10. Model eastward component of velocity for Case 2.3, incorporating faults and further weakened central Tibetan Plateau. Model faults are shown as thick black lines. Thin lines denote fault traces from the GEM Global Active Faults Database (Styron & Pagani, 2020).

Figure 11. (a) Maximum shear strain rate and (b) dilatation from the geodetically-derived velocity field (Figure 1, Wright et al., 2023). (c) Model-derived maximum shear strain rate from Case 2.3: incorporation of faults and additional weak central Tibetan Plateau. Arrow pairs show principal strain rates, with contraction shown in gray and extension shown in blue. (d) Dilatation strain rate from Case 2.3. \nst = 10^{-9}.\n
\n
-19-
Figure 12. Zoomed view of observed (blue arrows) and model (red arrows) GNSS velocities for the best-fit solution (i.e., Case 2.3) in Tian Shan and northwestern Tibetan Plateau (a), plateau interior (c), northeastern (e), and southeastern plateau (g). The associated residual vectors are shown as magenta arrows in (b), (d), (f), and (h). The individual RMS misfit values for each region are 3.3 mm/yr (b), 4.6 mm/yr (d), 2.4 mm/yr (f), and 3.8 mm/yr (h), respectively. The spatial extents of each panel are indicated in Figure 1b.
glected when averaged over many fault cycles. We assume the ductile deformation can be described by a non-linear (power-law) viscous rheology. Barr and Houseman (1996) introduce faults into a viscous medium by applying zero shear stress on the faults, although the actual shear stress on active faults is poorly constrained. In this study we describe the deformation field in terms of a viscous continuum with faults on which slip is resisted. The dimensional fault-resistance coefficient depends on the choice of Argand number (see Eqs. 4 and 8). In the context of this model, faults can be locked, stress-free, or support a traction that is proportional to the slip rate. Our results show that the best-fit model requires some resistance to slip on faults (Figure 6). Locked faults do not slip and thus they cannot localize strain unless they coincide with strength-contrast boundaries (e.g., the Kunlun and Xianshuihe Faults in Case 1.4, Figures 6i, 6j, 6l, and 6m). Free-slipping faults overestimate the observed velocity steps (e.g., Figures 6e, 6f, 6i, 6j, 6l, and 6m). Our preferred model uses a uniform scaled fault-resistance coefficient of $f_t = 0.4 \text{ MPa}\cdot\text{yr/mm}$ for all model faults subject to the choice of $Ar = 4$ (Case 2.3), although the velocity contrasts appear to be over-predicted across the Altn Tagh Fault and the Xianshuihe Fault. Applying relatively large resistance coefficients for the two faults (e.g., $f_t = 3.5 \text{ MPa}\cdot\text{yr/mm}$) can improve the fits locally (Figures 6e, 6f, 6l, 6m, and S4).

**4.2 Comparison with Previous Dynamic Models of the India-Eurasia Collision**

Table 2 shows a compilation of what existing dynamic models of the India-Eurasia collision predict in terms of the key tectonic deformation patterns observed. Our numerical experiments can intrinsically predict large-scale distributed deformation in the India-Eurasia collision zone. The best model (Case 2.3) explains all the key observations from geodesy listed in Table 1 (see Figures 6, 9, 10, 11, and 12). Whilst we are fitting all of the longer-term features, there remain strong features that we are not expecting to fit, as they relate to shorter-timescale earthquake-cycle type processes, such as elastic locking along the Himalayas. The laterally homogeneous viscous sheet model (England & Houseman, 1986) does not predict the E-W extension of the plateau or focused strain around faults, but lithospheric strength discontinuities cause strain concentration (Dayem et al., 2009; Molnar & Dayem, 2010; Lechmann et al., 2014; Bischoff & Flesch, 2019). Our model distribution of effective viscosity (Figure 13) is comparable to those determined by Flesch et al. (2001), Liu and Yang (2003), Copley and McKenzie (2007), and Deng and Tesauro (2016). Our results support the findings of a strong ($10^{24} \text{ Pa s}$, Figure 13) Tarim Basin and a weak ($\sim 10^{22} \text{ Pa s}$) Tian Shan (Neil & Houseman, 1997). The Tarim Basin appears to behave as a secondary rigid indenter and experiences little internal deformation, but transmits stress and gives rise to local crustal thickening in Tian Shan (Figure 11) (Molnar & Tapponnier, 1975; England & Houseman, 1985; Neil & Houseman, 1997; Huangfu et al., 2021). We find that a relatively weak ($10^{22}-10^{23} \text{ Pa s}$) high topographic region ($\sim 2,000 \text{ m}$) predicts the dilatation of the highest-elevation region of the Tibetan Plateau and convergence on the margins of the plateau especially in the northeastern plateau (Cases 1.3, 1.4, 2.2, and 2.3, Figures S1f, S1h, S11, and 11b). Thus the E-W extensional collapse of the plateau may be explained either by increases in surface elevation (Liu & Yang, 2003) and GPE arising from the thermal evolution of thickened continental lithosphere (England & Houseman, 1989), or by a relatively weak Tibetan lithosphere with an average effective viscosity of $10^{21}-10^{22} \text{ Pa s}$ (England & Molnar, 1997; Flesch et al., 2001; Liu & Yang, 2003; L. Chen et al., 2017). A weak ($\sim 10^{21} \text{ Pa s}$) central Tibetan Plateau bounded by the Dianzhong Block provides an explanation for the smooth, long-wavelength eastward velocity variation away from major faults (Cases 1.4 and 2.3, Figure 6a), consistent with the suggestion that the Dianzhong Block obstructs the lithospheric extrusion in the southeastern Tibetan Plateau (Han et al., 2022). The asymmetric eastward velocity gradient across the Tibetan Plateau is mainly due to the
asymmetry of the external boundary conditions (Flesch et al., 2001; Bischoff & Flesch, 2019).

Slip on major faults (Case 2, Figures S1i, S1k, and 11a) and/or lithospheric strength contrasts (Cases 1.3 and 1.4, Figures S1e and S1g, Lechmann et al., 2014; Bischoff & Flesch, 2019) can produce focused strain. The clockwise rotation of the Tarim block (e.g., Avouac & Tappin, 1993; Z.-K. Shen et al., 2001; Craig et al., 2012; J. Zhao et al., 2019) is enhanced by motion on the Altyn Tagh Fault (Case 2, Figures 7c, 8c, and 9c); this rotation was not evident in the experiments of Flesch et al. (2001), Liu and Yang (2003), Lechmann et al. (2014), and Bischoff and Flesch (2019) as they did not take account of relative motion on the fault. The clockwise rotation around the EHS was obtained by Bischoff and Flesch (2019) invoking a west-to-east decrease in upper crustal strength.

In our numerical simulations, allowing for local convergence on the Sagaing Fault and eastern boundary of the Indian Plate allows the displacement pattern around the EHS to be simulated, and is justified as an approximate characterization of subduction in the Myanmar region (Case 2, Figures 7c, 8c, and 9c, e.g., Steckler et al., 2008).

4.3 Active Faulting and Seismicity

Although the preferred model includes several lithospheric-scale faults on which fault-like displacements are explicitly represented, we also consider that continuous strain within the ductile regions must also be manifest in smaller-scale faulting of the uppermost brittle layer to allow a deformation that is conformable with the continuous strain occurring in the ductile layers beneath. To evaluate the style of faulting expected at any given location we consider the triaxial strain rate field as a sum of two double couples aligned with the principal horizontal strain-rate axes ($\dot{\varepsilon}_1$) and ($\dot{\varepsilon}_2$) (Houseman & England, 1986). The style of faulting that covers the spectrum from normal to strike-slip to
<table>
<thead>
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<th>Key observations</th>
<th>2-D modeling</th>
<th>3-D modeling</th>
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<tr>
<td></td>
<td>(Case 2.3)</td>
<td></td>
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<tr>
<td>Distributed deformation throughout the India-Eurasia collision zone</td>
<td>✓</td>
<td>✓</td>
</tr>
<tr>
<td>Dilatation of high plateau</td>
<td>✓</td>
<td>×</td>
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<tr>
<td>Contraction on the margins of plateau</td>
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<td>✓</td>
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<tr>
<td>Smooth, long-wavelength eastward velocity variation away from major faults</td>
<td>✓</td>
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<tr>
<td>Strain concentrations on major faults</td>
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<tr>
<td>Asymmetric eastward velocity gradient across the Tibetan Plateau</td>
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<td>Clockwise rotation around the EHS</td>
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<td>–</td>
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<tr>
<td>Clockwise rotation of the Tarim basin</td>
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Table 2. Comparison of Dynamic Models Predicting the Key Features of the India-Eurasia Collision
reverse faulting can then be described using the parameter $p$:

$$p = \frac{3}{4} + \frac{1}{\pi} \arctan\left(\frac{\dot{\varepsilon}_2}{\dot{\varepsilon}_1}\right)$$  \hspace{1cm} (9)

When $p$ is in the range $0 \leq p < 0.25$ reverse faulting (RR) is predicted in both principal directions. When $0.25 \leq p < 0.5$ reverse faulting plus subsidiary strike slip (RS) or strike slip plus subsidiary reverse faulting (SR) is predicted, with the transition between RS and SR taking place where $p = 0.375$. Pure strike-slip faulting occurs when $p = 0.5$ and then transitions from strike slip with subsidiary normal faulting (SN) to normal faulting with subsidiary strike slip (NS) and from NS to NN take place at $p = 0.625$ and $p = 0.75$, respectively (Houseman & England, 1986; Gordon & Houseman, 2015; England et al., 2016; Walters et al., 2017).

Figure 14 shows the comparison between the predicted distribution of active faulting and the observed earthquake focal mechanisms. The classification of the focal mechanism data was performed using FMC program according to the values of the P, T, and B Centroid Moment Tensor axes (Álvarez Gómez, 2019). The edges of the plateau are characterized primarily by compressional strain/reverse faulting (Figure 14a). Strike-slip faulting occurs everywhere in the region (Figure 14b). Normal faulting is predicted to dominate in the plateau interior, especially in the southern plateau (Figure 14c). These calculated styles of deformation are in agreement with the distribution of earthquake focal mechanisms (Figure 14d), implying that the faults within the seismogenic upper crust are taking up strain imposed by the ductile lithosphere. This consistency between model prediction and observation validates a key assumption of negligible vertical gradients of horizontal velocities for the TVS model of the India-Eurasia collision in which the deformation field is explained by the balance between gravitational buoyancy forces and stress caused by plate convergence, moderated by a viscous constitutive law.

The TVS approach averages the rheological parameters over the thickness of the lithosphere, and thereby ignores the depth variation of those rheological parameters. We therefore have not considered the class of models in which lower crustal flows at a different rate to the surface (e.g., Royden et al., 1997; Clark & Royden, 2000; Copley & McKenzie, 2007; Royden et al., 2008). For example, Copley and McKenzie (2007) invoked a gravitational flow with rigid base that explains the deformation along the Himalayas and the Indo-Burman Ranges. The vertical partitioning of lithospheric strength is still debated (e.g., Schmalholz et al., 2018; M. Wang et al., 2021). Despite this, our estimate of depth-averaged effective viscosity provides a first-order constraint on the vertical variations of lithospheric strength whatever the depth-dependence of the viscosity profile. The TVS method also treats the lithosphere as a purely viscous medium, as the elastic strain is not represented in the long-term geological record and may be ignored if the inter-seismic strain rate field is representative of the long-term strain (Barr & Houseman, 1996). The simplicity of the TVS approximation allows us to explore the rheology of the lithosphere and gain insights into the behavior of faults in a viscous continuum and the relationship between active faulting and seismicity.

Although relatively complex, our preferred model is necessarily simplified compared to reality, with assumptions like piece-wise constant $B'$ and constant $f'$. Further fine-tuning of these model parameters or adding additional complexity in boundary conditions might produce a more exact fit to data, particularly along the Himalayan arc and in the southeastern Tibetan Plateau, but would probably not change the broad conclusions reached here. However, possible lateral variations of GPE determined by the thermal evolution of the thickened lithosphere could mitigate the requirement for a very weak central Tibetan Plateau. Apparent misfits are likely controlled by the 3-D nature of collision which is not accounted for in the TVS model (Steckler et al., 2008; Artemieva et al., 2016).
Figure 14. Predicted distribution of fault types compared with observed earthquake focal mechanisms (magnitude ≥ 5.0) from the GCMT catalog (Dziewonski et al., 1981; Ekström et al., 2012). In the two-letter designations, N, S, R, refer to normal, strike-slip, and reverse faulting, with the first letter representing the dominant style of deformation. The $p = 0.5$ contours are shown as gray lines. Purple lines indicate the boundary of the calculation domain. Thick black lines are model faults. (a) Reverse-faulting earthquakes of the region. (b) Strike-slip-faulting earthquakes. (c) Normal-faulting earthquakes. (d) Percentage of earthquake focal mechanisms compared with calculated dominant styles of deformation.
5 Conclusions

We have shown that two-dimensional dynamic models based on a thin viscous shell formulation incorporating discontinuous displacement on major faults can explain the key observations of the India-Eurasia convergence as expressed in the new high-resolution Sentinel-1 InSAR as well as GNSS velocity fields. We conclude that:

(1) The balance between gravitational buoyancy-induced stress and viscous stress shapes the deformation field in the India-Asia collision zone; the preferred model fits the combined geodetic observations with an RMS misfit of 3.5 mm/yr and an Argand number of $\sim 4.0$.

(2) The observed dilatation strain rate field is explained by the inclusion of a relatively weak region of high topography ($\sim 2,000$ m) with a depth-averaged effective viscosity of $10^{22} - 10^{23}$ Pa s.

(3) A weak central Tibetan Plateau ($\sim 10^{21}$ Pa s) bounded by the Dianzhong Block replicates the smooth, long-wavelength eastward velocity variation away from major faults.

(4) Shear resistance to slip (0.4 MPa-yr/mm subject to the choice of $\text{Ar}=4$) on major faults allows strain concentration on those systems.

(5) Clockwise rotation around the EHS is produced by the model allowing for local convergence on the eastern boundary of the Indian Plate (7.5 MPa-yr/mm) and the Sagaing Fault (0.4 MPa-yr/mm), approximately representing subduction in the Myanmar region.

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Supporting Information for “The Dynamics of the India-Eurasia Collision: Faulted Viscous Continuum Models Constrained by High-Resolution Sentinel-1 InSAR and GNSS Velocities”
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Figure S1.  (a, c, e, g, i, k, m) Model-derived maximum shear strain rate from each case. Arrow pairs show principal strain rates, with contraction shown in gray and extension shown in blue. (b, d, f, h, j, l, n) Dilatation strain rates from the models.
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Figure S3. Test of Gaussian filtering width to capture the velocity gradient across a fault. Blue dashed line denotes a representative velocity profile from the dynamic model. Red dotted line represents an arctan-shape velocity profile for a fault with a slip rate of 10 mm/yr and a locking depth of 14 km. Solid lines show different filter widths applied. A width of 100 km (magenta line) best approximates the velocity gradient.
Figure S4. (a) Fits to observations from the calculation with relatively large fault-resistance coefficients ($f_t = 3.5 \text{ MPa} \cdot \text{yr/mm}$) applied to the Altyng Tagh Fault and the Xianshuihe Fault, and $f_t = 0.4 \text{ MPa} \cdot \text{yr/mm}$ for the rest of model faults. Other parameter settings are kept the same as Case 2.3. (b) Associated residual vectors.