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1 Dynamics of continental deformation in Asia

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5 **Abstract.** The relevance of plate tectonics concepts to the description of deformation
6 of large continental areas like Asia is subject to much debate. For some, the deforma-
7 tion of continents is better described by rigid motion of lithospheric blocks with strain
8 concentrated along narrow fault zones. For others, it is better described by viscous flow
9 of a continuously deforming solid in which faults play a minor role. Discriminating these
10 end-member hypothesis requires spatially dense measurements of surface strain rates cov-
11 ering the whole deforming area. Here, we revisit the issue of the forces and rheological
12 structure that control present-day deformation in Asia. We use the “thin-sheet” theory,
13 with deformation driven by the balance of boundary and buoyancy stresses acting on
14 a faulted lithosphere with laterally varying strength. Models are validated against a re-
15 cent, homogeneous, GPS velocity field that covers most of Asia. In the models, deforma-
16 tion in compressional areas (Himalayas, Tien Shan, Altay) is well reproduced with
17 strong coupling at the India/Eurasia plate contact, that allows for boundary forces to
18 transfer into Asia. Southeastward motions observed in north and south China however
19 require tensional, oceanward-directed stresses, possibly generated by gravitational po-
20 tential energy gradients across the Indonesian and Pacific subductions. Model and ob-
21 served strain rates show that a large part of Asia undergoes no resolvable strain, with
22 a kinematics apparently consistent with block- or plate-like motions. Internal strain, pos-
23 sibly continuous, is limited to high-elevation, mechanically weaker, areas. Lateral vari-
24 ations of lithospheric strength appear to control the style of deformation in Asia, with
25 a dynamics consistent with the thin-sheet physical framework.

1. Introduction

26 The success of plate tectonics is due, for a large part, to
27 its ability to correctly describe horizontal surface motions
28 for most of our planet by simple rotations of a limited
29 number of rigid plates. Indeed, geodetic measurements of
30 the relative motion of sites located far enough away from
31 plate boundaries show a remarkable agreement with the
32 theory [Robbins *et al.*, 1993; Argus and Heflin, 1995] and,
33 with only rare exceptions, the oceanic parts of plates do
34 not deform significantly. In the continents, however, the
35 relevance of plate tectonic concepts to describe horizontal
36 motions remains debated [e.g., Molnar *et al.*, 1973; Mol-
37 nar and Tapponnier, 1975; Thatcher, 2003; England and
38 Molnar, 2005]. Seismicity is diffuse and geologic struc-
39 tures show that deformation can affect broad areas, sug-
40 gesting more complex processes than in the oceans. For
41 some, deformation of continents is localized on a lim-
42 ited number of major faults bounding rigid lithospheric
43 blocks and is driven solely by stresses due to the motions
44 of neighboring plates. For others, deformation is perva-
45 sive and driven, for a significant part, by buoyancy forces
46 resulting from lateral variations of crustal thickness.

47 Conceptual models of continental deformation in Asia
48 follow this bimodal pattern. Edge-driven models (implic-
49 itly assuming plane-strain) argue that boundary stresses
50 due to the India-Eurasia collision are responsible for the

51 eastward extrusion of rigid lithospheric blocks bounded
52 by fast-slipping lithospheric-scale faults [e.g., *Tappon-*
53 *nier et al.*, 1982; *Peltzer and Saucier*, 1996]. *Peltzer*
54 *and Saucier* [1996] used these assumptions to numerically
55 simulate the deformation of Asia and found a good
56 fit to geological data and to the sparse geodetic obser-
57 vations available at the time. On the other hand, thin-
58 sheet models (implicitly assuming plane-stress) treat the
59 lithosphere as a continuous viscous medium where defor-
60 mation is accommodated by crustal thinning or thicken-
61 ing. The resulting spatial variations in crustal thickness
62 induce lateral variations in gravitational potential en-
63 ergy (GPE) that, in turn, contribute to the force balance
64 driving deformation [*Frank*, 1972; *Molnar and Tappon-*
65 *nier*, 1978; *Vilotte et al.*, 1982; *England and Houseman*,
66 1986; *Cobbold and Davy*, 1988; *Houseman and England*,
67 1986, 1993]. For instance, *England and Molnar* [1997a]
68 claim that a model in which horizontal gradients of the
69 deviatoric stress required to deform a thin viscous sheet
70 are balanced by horizontal GPE gradients explains first-
71 order active deformation features in and around Tibet.
72 Using a similar approach with the added constraint of a
73 sparse GPS data set, *Flesch et al.* [2001] argue that GPE
74 contributes to about 50% of the force balance driving
75 present-day deformation in Asia.

76 While the advances made in modeling continental
77 deformation in Asia during the past 20 years have
78 been impressive, these studies have suffered from lim-
79 ited quantitative data to compare theoretical predictions
80 against. Geodetic measurements are now providing a
81 wide-aperture image of present-day surface displacements
82 in Asia with precision of 1-2 mm/yr [e.g., *Wang et al.*,
83 2001; *Calais et al.*, 2006]. This, coupled with the physical
84 framework developed by *Houseman and England* [1986]
85 and later modified by *Bird* [1989], allows us to revisit the
86 issue of the forces and lithospheric rheology that control
87 present-day deformation in Asia.

2. Active Deformation in Asia

88 Active deformation in Asia has been extensively stud-
89 ied over the past 30 years from tectonic [e.g., *Tappon-*
90 *nier and Molnar*, 1979; *Burchfield and Royden*, 1991; *Zhang*
91 *et al.*, 1995], seismological [e.g., *Molnar et al.*, 1973; *Mol-*
92 *nar and Deng*, 1984], and geological or paleoseismological
93 observations [e.g., *Ritz et al.*, 1995, 2003; *Mériaux et al.*,
94 2004; *Lacassin et al.*, 2004]. It is well established that
95 current deformation is distributed over a broad area ex-
96 tending from the Himalayas in the south to the Baikal
97 rift to the north, and from the Pamir-Tien Shan to the
98 west to the peri-asiatic oceanic subductions to the east
99 (Figure 1). Seismicity is widespread and, although most
100 earthquakes occur at the Pacific subductions, a number
101 of large events have also struck the interior of the con-
102 tinent, such as four M_w 8.0-8.4 earthquakes in Mongolia
103 between 1905 and 1957 [*Khilko et al.*, 1985; *Okal*, 1977;
104 *Baljinnyam et al.*, 1993] or, more recently, a M_w 7.4 event
105 in the Russian Altay (September 2003, *Bourtchevskaia*
106 *et al.* [2005]).

107 The analysis of Landsat imagery in the late 1970's
108 [*Molnar and Tapponnier*, 1975; *Tapponnier and Mol-*
109 *nar*, 1977, 1979], together with geologic and seismolog-
110 ical data [e.g., *Molnar and Deng*, 1984; *Molnar et al.*,
111 1987], led to the idea that Quaternary deformation in
112 Asia was accommodated by slip on a limited number of
113 large faults bounding aseismic – therefore assumed to
114 be non-deforming – blocks (Tarim basin, South China,
115 North China, Sunda; Figure 1). The boundaries of these
116 blocks include large strike-slip faults (e.g., Altyn Tagh,

117 Karakorum, Kunlun, Xianshuihe, Jiali, Haiyuan, Gobi
118 Altay; Figure 1) as well as compressional ranges (e.g.,
119 Himalayas, Pamir-Tien Shan, Mongolian Altay, Gobi Al-
120 tay; Figure 1). In addition to these localized deformation
121 zones, two broad high elevation areas, the Tibetan and
122 Mongolian plateaus, characterize the long wavelength to-
123 pography of Asia, with average elevations of 5,000 and
124 2,500 m and crustal thicknesses up to 70 and 50 km, re-
125 spectively. The mechanism that led to such amounts of
126 crustal thickening over broad areas and their impact on
127 present-day deformation are still debated.

128 The past decade has seen a rapid increase in geodetic
129 results in Asia [e.g., *Abdrakhmatov et al.*, 1996; *Bilham*
130 *et al.*, 1997; *King et al.*, 1997; *Paul et al.*, 2001; *Calais*
131 *et al.*, 1998; *Heki et al.*, 1999; *Chen et al.*, 2000; *Shen*
132 *et al.*, 2000; *Kogan et al.*, 2000; *Calais and Amarjargal*,
133 2000; *Bendick et al.*, 2000; *Shen et al.*, 2001; *Wang et al.*,
134 2001; *Michel et al.*, 2001; *Bock et al.*, 2003; *Calais et al.*,
135 2003; *Wright et al.*, 2004; *Chen et al.*, 2004; *Wallace et al.*,
136 2004]. In some instances, these studies provide important
137 insight into the dynamics of continental deformation in
138 Asia. For instance, GPS and InSAR data show that the
139 central part of the Altyn Tagh fault accumulates strain
140 at a rate of 9 mm/yr [*Bendick et al.*, 2000; *Shen et al.*,
141 2001; *Wallace et al.*, 2004; *Wright et al.*, 2004], incon-
142 sistent with edge-driven block models that require slip
143 rates at least a factor of two larger [*Peltzer and Saucier*,
144 1996]. On the other hand, geodetic measurements of the
145 eastward velocity of south China at 8 to 10 mm/yr [e.g.,
146 *Wang et al.*, 2001] match block models and continuous
147 deformation models equally well [*Peltzer and Saucier*,
148 1996; *Molnar and Gibson*, 1996] but proved wrong early
149 models of extrusion that required at least 10-15 mm/yr
150 of eastward motion of south China [*Avouac and Tappon-*
151 *nier*, 1993]. At a continent-wide scale, *Flesch et al.* [2001]
152 showed that GPS data and Quaternary fault were con-
153 sistent with large parts of Asia undergoing little internal
154 strain. *England and Molnar* [2005], using similar data
155 but a different spatial resampling of the GPS velocities,
156 argue that continuous deformation, driven for a large part
157 by gravitational potential energy (GPE) gradients, dom-
158 inate. Therefore, even though it is now accepted that
159 only 20 to 30% of India-Eurasia convergence is accom-
160 modated by lateral extrusion of continental blocks [e.g.,
161 *Peltzer and Saucier*, 1996; *England and Molnar*, 1997b],
162 no consensus has yet been reached on the processes and
163 relative importance of the forces that drive present-day
164 continental deformation in Asia.

165 Here, we use a new GPS velocity field derived from
166 a geodetic combination of three solutions in a consistent
167 reference frame [*Calais et al.*, 2006] that provides hori-
168 zontal motions at 165 sites in Asia with reasonably even
169 station spacing (Figure 2). We refer the reader to *Calais*
170 *et al.* [2006] for a complete description of the data analysis
171 procedure and kinematic implications. This velocity field
172 shows (1) unresolvable strain rates ($< 3 \times 10^9$ /yr) over
173 a large part of Asia, with current motions well-described
174 by the rigid rotation of a limited number of microplates
175 (North China, South China, Tarim Basin, Sunda), and
176 (2) internal strain, possibly continuous, limited to high-
177 elevation areas. Many of the prominent features previ-
178 ously reported by local geodetic studies and predicted by
179 deformation models are present, in particular the NNE-
180 SSW shortening between India and the Tarim basin,
181 accommodated by the eastward motion of Tibet and
182 south China and the clockwise rotation of eastern Tibet
183 around the eastern Himalayan syntaxis, and the NNW-
184 SSE shortening across the Tien Shan at ~ 20 mm/yr
185 in the west, decreasing eastward to ~ 10 mm/yr. The

186 east- to southeastward velocities observed in Mongolia
 187 and North China are however not reproduced by most
 188 deformation models of Asia, whether extrusion- or thin-
 189 sheet-based. Also, GPS data show that the Baikal rift
 190 zone is currently opening at 4 ± 1 mm/yr [Calais *et al.*,
 191 1998], whereas an extrusion model predicts 0-1 mm/yr
 192 [Peltzer and Saucier, 1996]. Finally, GPS data in China
 193 show a low shortening rate across the eastern border of
 194 the Tibetan plateau [< 3 mm/year, Chen *et al.*, 2000;
 195 Shen *et al.*, 2000], but 5 to 11 mm/yr of far-field con-
 196 vergence between eastern Tibet and south China [Shen
 197 *et al.*, 2000; Wang *et al.*, 2001]. This result is at odds
 198 with extrusion models, in which the eastward motion of
 199 South China is driven by the extrusion of Tibet in re-
 200 sponse to India-Eurasia collision. As proposed by King
 201 *et al.* [1997], this absence of significant shortening across
 202 the eastern Tibetan margin might indicate that the Ti-
 203 betan crust is actually rotating clockwise around the east
 204 Himalayan syntaxis, in agreement with southward veloc-
 205 ities observed in southwestern China. In that interpre-
 206 tation, the eastward motion of south China is occurring
 207 independently of the motion of Tibet.

208 Discrepancies between previous models (with very few
 209 GPS data at their disposal) and recent GPS observations
 210 may arise from the choice of (1) boundary conditions, (2)
 211 sources of deviatoric stresses (boundary versus buoyancy
 212 forces, mantle tractions), or (3) lithospheric rheology. To
 213 investigate these issues, we model present-day deforma-
 214 tion in Asia using a finite element code that simulates the
 215 deformation of a faulted lithosphere on a spherical Earth.
 216 Deformation is driven by the balance of boundary stresses
 217 resulting from current plate motions and interplate cou-
 218 pling, and buoyancy stresses caused by horizontal gradi-
 219 ents of gravitational potential energy. These stresses act
 220 on a faulted lithosphere with a vertically-integrated rhe-
 221 ology consistent with laboratory rock experiments and
 222 regional heat flow data. Model outputs are horizontal
 223 surface velocities and fault slip rates, which we compare
 224 with corresponding observations from GPS and Quater-
 225 nary geology.

3. Deformation Model

3.1. Model Assumptions

226 We use the finite element code “SHELLS”, which has
 227 previously been applied to modeling active deformation
 228 in intra- and inter-plate settings in Asia, Alaska, Califor-
 229 nia, Mediterranean, and New Zealand [Bird and Baum-
 230 gardner, 1984; Bird and Kong, 1994; Bird, 1996; Kong
 231 and Bird, 1996; Jiménez-Munt *et al.*, 2001, 2003; Liu and
 232 Bird, 2002a]. We refer the reader to Bird [1989], Kong
 233 and Bird [1995] and Bird [1999] for a complete descrip-
 234 tion of its physical basis and assumptions. We summarize
 235 hereafter its major characteristics.

236 SHELLS approximates the lithosphere as a spherical
 237 shell of variable thickness, where the horizontal compo-
 238 nents of the momentum equation are radially integrated
 239 and velocities are constant with depth (“thin plate”
 240 approximation). Calculations are iterated until quasi-
 241 steady state is reached, using time-invariant boundary
 242 conditions, so that elasticity contributes negligibly to the
 243 strain rate. SHELLS assumes incompressibility, which
 244 is consistent with neglecting elastic strain. It uses an
 245 anelastic rheology with frictional sliding on faults in the
 246 upper crust and upper part of the lithospheric mantle,
 247 and a non-Newtonian, thermally activated dislocation
 248 creep for the lower crust and the lower part of the litho-
 249 spheric mantle. SHELLS assumes two sets of constant
 250 thermal parameters for the crust and the mantle and ac-

251 counts for spatial variations of surface heat flow, which is
252 interpolated from actual measurements. Hence, SHELLS
253 simulates a layered continental lithosphere similar to the
254 classical view derived from rock physics experiments and
255 seismological observations, where a weak (ductile) lower
256 crust overlies a strong upper mantle [*Brace and Kohlstedt*,
257 1980; *Chen and Molnar*, 1983; *Kirby and Kronenberg*,
258 1987]. The strength of the lithosphere is determined
259 by 3D numerical integrals that take into account spa-
260 tial variations of geotherm, crust and mantle properties,
261 and strain rates. SHELLS does not account for flexural
262 strength (no vertical shear traction on vertical planes),
263 which implies that vertical normal stress is lithostatic at
264 all points of the model. The lithosphere is divided into
265 triangular elements and the horizontal components of ve-
266 locities are solved at all the nodes. These velocities are
267 averaged over longer time than that of the seismic cy-
268 cle and the computation ignores all accelerations except
269 gravity.

3.2. Grid Geometry and Boundary Conditions

270 We built a finite element grid made of 2313 nodes
271 and 3336 triangular elements (Figure 3). We densified
272 it in the Mongolia-Baikal area for the purpose of a de-
273 tailed comparison with recent geodetic results in this area
274 [*Calais et al.*, 2003]. Since we use the Eurasian plate as a
275 reference, we impose zero-displacement along the north-
276 ern and most of the western borders of the model. We
277 allow for north-south displacements along the southern-
278 most part of its western border, corresponding to the
279 Zagros orogeny (Figure 3). The other grid borders fol-
280 low major plate boundaries, along which we use velocity
281 boundary conditions taken from the REVEL plate kine-
282 matic model (*Sella et al.*, 2002). REVEL velocities agree
283 within 1-2 mm/yr with those used here for sites on plate
284 interiors (in particular in India), and therefore provide
285 boundary conditions consistent with the GPS velocity
286 field used in this study.

287 The Arabia-Eurasia plate boundary follows the main
288 thrust zone of the Zagros orogeny and the Makran
289 oceanic subduction. The India-Eurasia plate boundary
290 follows the left-lateral strike-slip fault zones of Pakistan,
291 the Himalayan frontal thrust, the right-lateral strike-
292 slip fault system of Burma (*Sagaing*)-Andaman. The
293 Australian-Eurasia boundary follows the oceanic subduc-
294 tion of the Australia plate under Sumatra and the In-
295 donesian arc, with the Sorong and Palu strike-slip faults
296 and the north Sulawesi and Moluccu subductions ac-
297 counted for. The Philippines-Eurasia plate boundary
298 follows the subduction of the Philippines Sea plate un-
299 der the Philippines Archipelago and includes the Philip-
300 pines strike-slip fault and the subduction reversal of the
301 Manilla trench, in continuation with the Taiwan orogeny
302 and further north with the Ryu-Kyu and Nankai sub-
303 ductions. The Pacific-Eurasia plate boundary follows
304 the Japan, Kuriles, and Kamchatka subductions and ac-
305 counts for the Main Seismic Line in Japan and the Kuriles
306 strike-slip fault zones. The Eurasia-North America plate
307 boundary follows the Nansen ridge.

3.3. Faults

308 We used the trace of major active faults in Asia as
309 reported by *Sherman* [1978], *Houdry* [1994], *Levi et al.*
310 [1995], *Moore et al.* [1997] and *Agar and Klitgord* [1995]
311 for the Altay-Sayan-Baikal-Stanovoy area, from *Schlupp*
312 [1996] for Mongolia, *Tapponnier and Molnar* [1977]; *Tap-*
313 *ponnier et al.* [1982], and from *Replumaz* [1999] for the
314 rest of Asia (Figure 3). We assigned a single dip angle

315 value for each type of fault: 65° for normal faults, 30° for
 316 thrust faults, 90° for strike-slip faults, and 45° for faults
 317 whose dip and sense of slip is uncertain. The dip an-
 318 gle for the oceanic and continental subductions at plate
 319 boundaries is 23°. The faults are free to slip in any di-
 320 rection, regardless of their dip angles, except for faults
 321 dipping 90°, which are forced to move in a strike-slip
 322 sense. Slip on faults is continuous and controlled by the
 323 deviatoric stress in their vicinity and by a single fault
 324 friction coefficient for the entire model. Continuous ele-
 325 ments in the model are distinguished from faults by their
 326 higher internal friction coefficient.

327 In order to simulate mechanical coupling between
 328 plates, we limit shear traction at convergent boundaries
 329 to a maximum value above which slip occurs. Imposing a
 330 high maximum shear traction results in large stresses to
 331 be transferred to the continent (high coupling). The op-
 332 posite is true when imposing a low maximum shear trac-
 333 tion (low coupling). The model accounts for two maxi-
 334 mum shear traction values, one for continental subduc-
 335 tions (India, Arabia), one for oceanic subductions (Aus-
 336 tralia, Philippines, Pacific).

3.4. Lithospheric Structure

337 We extracted topography and bathymetry from the
 338 ETOPO5 global database, resampled at the model nodes
 339 using a piecewise-planar interpolation. We used heat flow
 340 data from the worldwide database compiled by *Pollack*
 341 *et al.* [1993], and from *Lysak* [1992] for the Mongolia-
 342 Baikal area (Figure 4), also resampled at the model
 343 nodes using a piecewise-planar interpolation. Heat flow
 344 in continental Asia ranges between 50 and 80 mW/m²,
 345 with most values comprised between 60 and 70 mW/m²,
 346 except for two positive anomalies in the Hovsgol area
 347 (northern Mongolia) and the northern part of the Baikal
 348 rift zone, where heat flow reaches 100 and 120 mW/m²,
 349 respectively.

350 Once elevation and heat flow values are attributed to
 351 each grid node, crust and lithospheric mantle thicknesses
 352 are computed iteratively until isostasy is established
 353 and a given temperature (typically 1200 or 1300° C) is
 354 reached at the base of the lithosphere. Hence, the re-
 355 sulting lithospheric structure (and consequently its inte-
 356 grated strength) depends on thermal parameters (ther-
 357 mal conductivity, radioactive heat production, and vol-
 358 umetric thermal expansion coefficient for both the crust
 359 and the mantle) and on the crust and mantle density. A
 360 broad range of thermal parameters have been proposed
 361 in the literature for continental lithosphere. We chose to
 362 test two different parameter sets, one corresponding to
 363 a weak lithosphere (*Bird* [1998], *Liu and Bird* [2002b]),
 364 the other corresponding to a strong one (*Kong and Bird*
 365 [1996]), as summarized in Table 1.

3.5. Constitutive Laws

366 The model considers two constitutive laws, a frictional
 367 sliding for the upper crust and the upper part of the litho-
 368 spheric mantle, a non-Newtonian thermally activated dis-
 369 location creep law for the lower crust and the lower part
 370 of the lithospheric mantle, respectively given by:

$$371 \quad \sigma_s = \mu(-\sigma_n - P_p) \quad (1)$$

372 and

$$373 \quad \sigma_s = A\dot{\epsilon}^{1/n} e^{\left(\frac{B+Cz}{T}\right)} \quad (2)$$

374 where μ is the friction coefficient, P_p the hydrostatic pore

375 pressure, σ_s and σ_n the shear and normal stresses, $\dot{\epsilon}$ the
 376 strain rate, n the power-law exponent, T the absolute
 377 temperature, z the depth, and A , B , and C three rheo-
 378 logical constants, distinct for the crust and for the litho-
 379 spheric mantle. Values for A , B , and C are taken from
 380 *Bird and Kong* [1994] and *Kirby* [1983], and are sum-
 381 marized in Table 2. Laboratory-derived power-law ex-
 382 ponents for lower crustal and upper mantle rocks usually
 383 range from 2.5 to 4 [*Kirby and Kronenberg*, 1987]; we used
 384 $n = 3$. Friction coefficients for faulted and non-faulted
 385 elements will be derived from a parameter space search
 386 constrained by the observed GPS velocities (see below).
 387 The set of parameters chosen here, combined with the
 388 rheological laws given above, result in a factor of 2 differ-
 389 ence in integrated strength between the weak and strong
 390 lithosphere models (Figure 5).

3.6. Crustal and Lithospheric Thicknesses

391 Given a set of thermal and rheological parameters,
 392 present-day topography, and surface heat flow, crustal
 393 and lithospheric thicknesses are calculated assuming
 394 isostasy and a steady state thermal regime. Resulting
 395 lithospheric and crustal thicknesses (Figure 6) match well
 396 estimates from geophysical data over most of Asia [*Vil-
 397 laseñor et al.*, 2001; *Li and Mooney*, 1998; *Mooney et al.*,
 398 1998; *Lebedev and Nolet*, 2003; *Artemieva and Mooney*,
 399 2001].

400 Model crustal thickness ranges between 30 and 40 km
 401 for most of continental Asia, but reaches 50 and 75 km
 402 under the Mongolian and Tibetan plateaus, respectively
 403 (Figure 6). It is generally less for the weaker than for
 404 the stronger lithosphere tested here. Crustal thickness
 405 is better reproduced with a weaker lithosphere for moun-
 406 tain areas and high elevation plateaus, where the stronger
 407 lithosphere results in slightly overestimated values. The
 408 opposite is true for oceanic and cratonic areas, where the
 409 stronger lithosphere matches observed crustal thicknesses
 410 slightly better than the weaker one.

411 Model lithospheric thickness varies from 100 to 160 km
 412 for the stronger lithosphere tested here (assuming a basal
 413 temperature of 1300° C) and match values estimated by
 414 *Artemieva and Mooney* [2001] from observational data,
 415 except under the Siberian craton where model values are
 416 underestimated by about 20 km. For the case of a weaker
 417 lithosphere, model thicknesses (assuming a basal temper-
 418 ature of 1200° C) are usually underestimated by about
 419 40 to 50 km compared to observations, except under Ti-
 420 bet where they match *Artemieva and Mooney* [2001]’s re-
 421 sults well. Uncertainties in the estimation of lithospheric
 422 thickness from observational data are however large (up
 423 to 50 km) because of uncertainties in thermal parameters,
 424 temperature at the base of the lithosphere, and surface
 425 heat flow. We found that variations in lithospheric thick-
 426 ness of that order have a negligible effect on its integrated
 427 strength, which is primarily sensitive to crustal thickness
 428 in the models.

4. Best-fit Rheological Parameters

4.1. Parameters Tested

429 In a first step, we seek to determine appropriate values
 430 for the internal and fault friction coefficients and for the
 431 maximum shear tractions at oceanic and continental sub-
 432 ductions. To do so, we run a series of models that system-
 433 atically explore this 4-parameter space and score them
 434 using the root-mean-square (*RMS*) misfit of model to
 435 observed horizontal velocities. Scoring only includes sites
 436 with velocity uncertainties less than 1.5 mm/yr (95% con-

437 fidence). We perform this grid search for both the weak
 438 and strong lithospheres tested here and retain the set of
 439 parameters that results in the smallest *RMS*.

440 We tested internal friction coefficient values (f_i) in the
 441 0.5-0.9 range, according to standard values derived from
 442 rock physics experiments [e.g., *Byerlee, 1978*]. A number
 443 of mechanical and thermal approaches [e.g., *Cattin, 1997*;
 444 *Cattin et al., 1997*] or modeling results [e.g., *Bird and*
 445 *Kong, 1994*; *Bird, 1998*; *Kong and Bird, 1996*; *Jiménez-*
 446 *Munt et al., 2001*; *Jiménez-Munt and Negredo, 2003*; *Liu*
 447 *and Bird, 2002a*] find fault friction coefficients (f_f) signif-
 448 icantly smaller than internal friction, with values usually
 449 less than 0.2. We therefore tested fault friction coefficient
 450 in the 0.04-0.20 range.

451 In an early deformation model of Asia, *Bird [1978]*
 452 used maximum shear traction on the Indian continental
 453 subduction (T_c) in the 20-30 MPa range, a value later
 454 amended to 15 MPa by *Kong and Bird [1996]* and *Lesne*
 455 *[1999]*. The same authors proposed a range of possible
 456 values for the maximum shear tractions at oceanic sub-
 457 ductions (T_o). On the basis of these previous studies, we
 458 tested values ranging from 15 to 30 MPa for T_c and from
 459 0.2 to 6 MPa for T_o .

4.2. Results

460 The *RMS* misfit between modeled and observed ve-
 461 locities range from 4.2 to 16.8 mm/yr (Figure 7). We
 462 find the smallest *RMS* for a maximum shear traction of
 463 20 MPa or higher at continental subductions and 4 MPa
 464 or higher at oceanic subductions, regardless of the strong
 465 versus weak lithosphere tested here. The best-fit value
 466 ranges from 0.7 to 0.9 for the internal friction coefficient
 467 and from 0.04 to 0.10 for the fault friction coefficient.

468 The lowest *RMS* for the strong lithosphere case is
 469 4.5 mm/yr, slightly higher than that obtained with the
 470 weaker lithosphere (Figure 7). Moreover, the parameter
 471 search for the strong lithosphere case does not show a
 472 clear minimum within the bounds imposed in the search.
 473 Predicted surface velocities using the best-fit parameter
 474 set for the strong lithosphere case are significantly faster
 475 than for the weaker lithosphere, by 10 to 75% in Tibet,
 476 100 to 180% in north China, Mongolia and Baikal, and
 477 50 to 100% in east and south China. In southeast Asia,
 478 both models predict velocities in good agreement with
 479 observed GPS velocities.

480 We find the best-fit set of parameters ($RMS =$
 481 4.2 mm/yr) using the weaker lithosphere with $f_i = 0.8$,
 482 $f_f = 0.06$, $T_c = 20$ MPa, and $T_o = 4$ MPa. Figures 8A
 483 (model and observed GPS velocities) and 8B (residual
 484 velocities) show predicted surface velocities derived from
 485 this set of parameters in good agreement with the ob-
 486 served ones, both in direction and magnitude, for Tibet,
 487 South China, Tien Shan, and Southeast Asia. Model ex-
 488 tension rate across the Baikal rift zone is 1.5-3 mm/yr,
 489 slightly lower than GPS observations. Observed velocity
 490 directions in the Altay, their clockwise rotation across the
 491 Gobi-Altay and central Mongolia, and the E- to SE-ward
 492 directions in North China are well reproduced. However,
 493 model velocities are overestimated in the Gobi desert
 494 (~ 9 mm/yr against 4 mm/yr observed), central Mongo-
 495 lia (6-7 mm/yr against 3-4 mm/yr observed) and north
 496 China (6-7 mm/yr against 3-5 mm/yr observed), and the
 497 model does not reproduce the rapid clockwise rotation
 498 observed around the eastern Himalayan syntaxis in the
 499 Yunnan. The model fit to the observed GPS velocities
 500 may possibly be improved by allowing for regional vari-
 501 ations in interplate coupling and/or friction coefficients,
 502 which is beyond the scope of this study.

5. Testing the Force Balance

503 In a second step, we seek to quantify the relative im-
504 portance of (1) boundary forces resulting from the rela-
505 tive motion of neighboring plates and interplate coupling
506 (using a maximum shear traction as a proxy for stress
507 transfer at subductions), and (2) buoyancy forces result-
508 ing from GPE gradients. We use the thermal and consti-
509 tutive parameters described above and run a series of
510 experiments in which the contribution of GPE gradients
511 and relative plate motions are progressively added to the
512 models.

513 In a first model (Model 1, Table 3), we set the maxi-
514 mum shear traction at all convergent plate boundaries
515 to zero (including the India/Eurasia boundary along the
516 Himalayan front), thereby imposing free slip (or no me-
517 chanical coupling) between Asia and the adjacent plates.
518 Boundary conditions along the northern and western
519 sides of the model are fixed to zero displacement, as de-
520 scribed above. Deformation in this model is therefore
521 driven by buoyancy forces only. The resulting surface
522 velocity field (Figure 9) shows very fast south-directed
523 velocities in the Himalayas and southern Tibet and fast
524 southeast-directed velocities in southeast Asia that obvi-
525 ously do not match the observed GPS velocities in mag-
526 nitude or direction. Shortening in the Tien Shan and the
527 Mongolian Altay (western Mongolia) are not reproduced
528 either. Predicted velocity magnitudes in the rest of the
529 model are also overestimated, by about a factor of 5. Pre-
530 dicted extension across the Baikal rift zone is 10 mm/yr,
531 about 3 times faster than observed.

532 This first experiment shows, as expected, that buoy-
533 ancy forces alone are not sufficient to reproduce the major
534 compressional structures of Asia (Himalayas, Tien Shan,
535 Altay). The Himalayas and most of Tibet collapse south-
536 ward in the absence of a resisting boundary force along
537 the India/Eurasia continental subduction. This south-
538 directed collapse is a result of larger elevation differences
539 (and therefore GPE gradients) along the southern border
540 Tibet (India) than along its northern border with the
541 Tarim basin and Mongolian plateau. In spite of the very
542 large misfits, we note that this model driven by buoy-
543 ancy forces only reproduces well the SE-ward velocity
544 directions observed in north China, central and eastern
545 Mongolia, and south China, as well as the clockwise ro-
546 tation observed in GPS velocities in the vicinity of the
547 Gobi Altay.

548 In a second set of experiments, we add the contribu-
549 tion of boundary forces. First, we establish a resisting
550 force at the India/Eurasia and Arabia/Eurasia continen-
551 tal subductions by imposing a maximum shear traction
552 of 20 MPa along these boundaries, the value found in the
553 parameter search described above. We set the velocity of
554 the Indian and Arabian plates to zero. As previously, we
555 impose zero maximum shear traction at the oceanic sub-
556 ductions and keep the western and northern boundaries
557 of the model confined (Model 2, Table 3). The result-
558 ing model velocities (Figure 10) are close to zero in the
559 Himalayas and southern Tibet, as the collapse of these
560 structures found in the previous models is now resisted by
561 the strength of the India/Eurasia plate contact. Model
562 velocities show shortening along the northern edge of Ti-
563 bet, in the Tien Shan, and in the Altay and Gobi Altay
564 mountains, but at slower rates than observed. In the
565 rest of Asia, the velocity pattern is similar to the first
566 model described above, but generally larger (by a factor
567 of 2 to 3 from northwest Tibet to the Siberian platform,
568 1.2 in eastern Mongolia, Baikal and north China, 1.2 to
569 1.4 in southeast Asia). Velocities in south China remain

570 unchanged.

571 This experiment illustrates how coupling at the In-
572 dia/Eurasia plate boundary balances the effect of buoy-
573 ancy forces in Tibet and allows for compressional stresses
574 to be transferred into Asia. Interestingly, India-Eurasia
575 convergence is not necessary to reproduce the observed
576 east- and southeast-ward motions in most of north and
577 east Asia, provided that appropriate buoyancy forces are
578 acting along the eastern and southeastern sides of the
579 model.

580 Second, we apply the India/Eurasia relative plate ve-
581 locity at the India/Eurasia plate boundary along the Hi-
582 malayan front, and the Arabia/Eurasia relative plate ve-
583 locity at the Arabia/Eurasia plate boundary. All other
584 parameters are kept similar to the previous experiment
585 (Model 3, Table 3). Model velocities (Figure 11) are now
586 close to the observed ones in the Himalayas and southern
587 Tibet, and show shortening in the Tien Shan and Altay
588 mountains at a rate close to the GPS observations. Model
589 velocities are overall larger than in the previous model,
590 by a factor of 2 in the western half of Asia, 1.5 in Ti-
591 bet, Mongolia, Baikal and north China, and 1.2 in south
592 China and southeast Asia. Plate convergence and high
593 mechanical coupling between India and Eurasia therefore
594 allow us to reproduce the compressional strain observed
595 in the Himalayas, Tien Shan, and Mongolia Altay. How-
596 ever, velocities in most of north and east Asia are over-
597 estimated (up to 10 times in northern Asia), velocities
598 in Yunnan do not show the observed clockwise rotation
599 around the eastern Himalayan syntaxis, and velocities in
600 southeast Asia are not well reproduced.

601 Third, we establish a resisting force at the oceanic sub-
602 ductions in east and southeast Asia by imposing a max-
603 imum shear traction of 4 MPa, the value found in the
604 parameter search described above. We impose velocities
605 of the Australian, Philippines and Pacific plates to be
606 zero in all direction along their boundary with Eurasia.
607 All other model parameters are kept similar to the previ-
608 ous case (Model 4, Table 3). Compared to the previous
609 experiment, model velocities (Figure 12) decrease signifi-
610 cantly over the entire domain – as expected from adding
611 a resisting force along the eastern border of the model –
612 in particular in southeast Asia where they drop from 50-
613 55 mm/yr to 0-5 mm/yr. In the rest of Asia, velocities
614 decrease by a factor of 6 (south China) to 1.2 (Tibet)
615 but remain significantly faster than observed. Model ve-
616 locities are now directed eastward to east-northeastward,
617 instead of southeastward as found in the previous exper-
618 iments or in the GPS data.

619 Finally, we restore actual relative plate motions along
620 the oceanic subductions in east and southeast Asia, and
621 between North America and Eurasia at the northeastern
622 edge of the model. All other parameters are kept similar
623 to the previous case. This experiment therefore uses the
624 same parameters as the best-fit model described in sec-
625 tion 4.2 (Figure 8; Model REF, Table 3). Compared to
626 the previous experiment, model velocities decrease over
627 most of Asia (by a factor of 1.2 to 1.5 in the Tien Shan
628 and south China, and 1.5 to 2 in Mongolia-Baikal and
629 north China) to reach magnitudes that are now consis-
630 tent with the observations. Model velocity directions
631 are however systematically rotated counterclockwise by
632 10-15° compared to the observations in north and south
633 China. The fit is however very good in Southeast Asia,
634 both in magnitude and direction, although the observed
635 clockwise rotation around the eastern Himalayan syntaxis
636 is still not reproduced.

6. Discussion

6.1. Role of Buoyancy Forces

637 This series of experiments illustrates the interplay be-
638 tween buoyancy and boundary forces in driving present-
639 day deformation in Asia. We find that buoyancy forces
640 are significant overall and may drive, for a large part,
641 the east to southeastward motions observed in south
642 and north China and in central Mongolia. In our ex-
643 periments, these buoyancy forces need to be resisted by
644 compressional boundary forces along the oceanic sub-
645 ductions that bound Asia to the east and southeast in
646 order to match the observed velocities. These resisting
647 boundary forces result from the convergence of adjacent
648 oceanic plates toward Eurasia and interplate coupling at
649 the oceanic subductions. However, buoyancy forces alone
650 do not explain present-day deformation in most of the
651 western half of the domain, where N-S shortening dom-
652 inates (Himalayas, Tibet, Tien Shan, Altay). Deforma-
653 tion in these areas is well reproduced with strong cou-
654 pling at the India/Eurasia plate contact, that allows for
655 compressional boundary stresses to transfer into Asia.

656 *Flesch et al.* [2001] also underlined the importance
657 of buoyancy forces in driving deformation in Asia. In
658 their models, buoyancy forces result only from GPE gra-
659 dients across continental Asia. In order to further in-
660 vestigate the origin of the buoyancy forces acting in our
661 models, we show an additional experiment where we set
662 the bathymetry and heat flow in oceanic domains to -
663 100 m and 55 mW/m², respectively, thereby removing
664 GPE gradients – and the resulting buoyancy forces –
665 across oceanic margins. All other parameters are kept
666 identical to the reference experiment (Model 5, table 3).
667 We find that model velocities are similar to the refer-
668 ence model in the western half of Asia, whereas east-
669 ward motions in south and north China and SE Asia are
670 not reproduced anymore (Figure 13). We conclude that
671 the eastward motions obtained in our reference model in
672 eastern Asia are mostly driven by buoyancy forces orig-
673 inating at the eastern and southeastern oceanic margins
674 of Asia, where GPE gradients generate oceanward ten-
675 sional stresses that tend to pull east and southeast Asia
676 towards the subductions.

677 This may however be an oversimplification of the pro-
678 cess actually at work, since our models do not simulate
679 the actual dynamics of subduction zones. For instance,
680 we do not account for slab pull or for corner-flow be-
681 tween the upper plate lithosphere and the subducting
682 slab. Also, the maximum shear stress imposed at subduc-
683 tions in the models is a simple proxy for stress transfer
684 rather than a true physical representation of mechanical
685 coupling. However, it remains that the east- and south-
686 eastward motions observed in North China, South China,
687 and SE Asia are reproduced only if tensional, ocean-
688 ward stresses are applied to the southeastern and eastern
689 boundaries of the domain.

6.2. Could Boundary Forces Suffice?

690 In one of the first experiments on the dynamics of de-
691 formation in Asia, *Peltzer and Tapponnier* [1988] suc-
692 cessfully reproduced some of the geological observables
693 in Asia, using India-Eurasia collision as the only driv-
694 ing force, while the eastern and southeastern boundaries
695 of the model were left unconfined. *Peltzer and Saucier*
696 [1996] further quantified this result numerically assum-
697 ing an elastic, faulted, lithosphere and, again, India-Eurasia
698 collision as the only driving force. Both models predict
699 that strike-slip faults in Asia slip at fast rates, up to
700 20-30 mm/yr for the Altyn Tagh fault. Here, we test

701 whether a model driven by India-Eurasia collision alone
702 can explain the GPS current observations.

703 To simulate experimental conditions as close as possi-
704 ble to these early edge-driven models, we cancel the effect
705 of buoyancy forces by setting constant elevation (910 m)
706 and heat flow (65 mW/m^2) over the entire model. To
707 simulate a unconfined eastern boundary, we treat the
708 southeast and east Asia subductions as free displacement
709 boundaries, allowing slip in any direction. We decrease
710 the fault friction coefficient to 0.01 to enhance slip on
711 faults. To simulate the indenting mechanism of India,
712 we treat the India-Eurasia collision as a displacement
713 fixed boundary and impose the reference model velocities
714 along the Himalayan frontal thrust as boundary condi-
715 tions. Boundary conditions along the northern and west-
716 ern sides of the model are kept the same as the reference
717 model (section 3.2). Figure 14 shows that this experimen-
718 tal setup results in horizontal velocities that are in fair
719 agreement with the observed ones, with an *RMS* misfit
720 of 4.7 mm/yr. We also find that model velocities are sim-
721 ilar to those obtained in the reference model (Figure 8A)
722 and to those predicted by *Peltzer and Saucier* [1996].
723 Predicted fault slip rates in this experiment are higher
724 than in the reference model for the Karakorum and Altyn
725 Tagh faults, the two fastest slipping faults in the *Peltzer*
726 *and Saucier* [1996] model, but do not exceed 8 mm/yr
727 (Figure 17). Slip rates on other faults are lower than in
728 the reference model and than the observed Quaternary
729 rates (Figure 17). Shortening across the Himalayas, Tien
730 Shan, and Altay is also twice as small as the GPS ob-
731 servations or the reference model. Predicted extension
732 across the Baikal rift zone is also smaller than observed
733 (0.5-1.5 mm/yr against 3-4 mm/yr) and model veloci-
734 ties on the Khazak and Siberian plateforms (4-7 mm/yr)
735 differ significantly from the observations or the reference
736 model (0-3 mm/yr).

737 In spite of these differences, one may argue that a
738 model in which the deformation in Asia is forced only by
739 a velocity condition applied at the boundary between In-
740 dia and Eurasia provides a reasonable fit to current GPS
741 data. This however assumes that buoyancy forces in and
742 around Asia and boundary forces along the Indonesian
743 and Pacific subductions have a negligible contribution
744 to the force balance. Our experiment, as well as other dy-
745 namic models [e.g., *England and Molnar*, 1997a; *Flesch*
746 *et al.*, 2001; *England and Molnar*, 2005], however show
747 that GPE gradients in Asia generate buoyancy forces
748 comparable to boundary forces in their effect on surface
749 deformation. Also, large earthquakes at oceanic subduc-
750 tions in east and southeast Asia require some degree of
751 inter-plate coupling, inconsistent with the free boundary
752 assumption used in the model above. The model setup
753 used here is therefore likely to be missing some first-order
754 geophysical processes. We thus do not favor an inter-
755 pretation of the dynamics of the deformation based on
756 boundary forces alone.

6.3. Pervasive versus Localized Strain

757 The question of the forces driving continental deformation
758 in Asia is typically merged to that of strain distribu-
759 tion, with edge-driven models arguing for localized strain
760 on faults, while thin sheet models argue for continuous
761 deformation. Since the approach used here allows for slip
762 on faults and lateral variations of lithospheric strength, it
763 is useful to investigate the spatial distribution of strain in
764 our best-fit model, recalling that it includes both bound-
765 ary stresses and buoyancy stresses.

766 Figure 15 shows that model strain rates are not signif-
767 icant (*i.e.*, less than 3×10^{-9} /yr, the average precision of

768 the GPS measurements used here) over a very large part
 769 of Asia (most of North China, South China, and Sunda).
 770 They are significant in the Himalayas, central and eastern
 771 Tibet, the Pamir-Tien Shan and Altay ranges, Western
 772 Mongolia, and the Baikal rift zone. This result matches
 773 well the spatial distribution of strain rates derived di-
 774 rectly from GPS observations [Calais *et al.*, 2006]. It is
 775 also consistent with denser GPS measurements in Tibet
 776 [Chen *et al.*, 2004], which show that about 50% of the
 777 observed velocity field is explained by distributed defor-
 778 mation across the Tibetan plateau, the Qaidam basin,
 779 and the Qilian Shan, while the remaining 50% is accom-
 780 modated by slip on a limited number of faults. The same
 781 GPS measurements show a combination of east-west ex-
 782 tension and north-south compression in Tibet, which is
 783 well reproduced in our best-fit model (Figure 15).

784 From the best-fit model results, validated against the
 785 GPS observations, we calculate the effective lithospheric
 786 viscosity as the vertically averaged stress over strain-rate.
 787 The map of effective viscosity (Figure 16) shows lateral
 788 variations from $3 \times 10^{21} - 7 \times 10^{22}$ Pa s in Tibet and other
 789 high elevation areas, to $\sim 10^{24}$ Pa s in low elevation and
 790 cratonic areas. These effective viscosity estimations are
 791 consistent with previous results from Flesch *et al.* [2001],
 792 who also find lateral variations of vertically averaged ef-
 793 fective viscosity by up to 3 orders of magnitude between
 794 Tibet and its surroundings.

6.4. Fault Slip Rates

795 In addition to surface velocities, we calculate model
 796 slip rates on major active faults, that we compare to
 797 Holocene estimates from geologic data and to present-day
 798 estimates from geodetic data (Figure 17). We find that
 799 the predicted sense of motion on all faults matches geo-
 800 logic observations. Model slip rates match Holocene and
 801 geodetic rates well in the Baikal rift zone [San'kov *et al.*,
 802 2002], the Gobi Altay [Ritz *et al.*, 1995, 2003; Prentice
 803 *et al.*, 2002], and the Tien Shan [Avouac, 1991; Molnar
 804 and Deng, 1984; Abdrakhmatov *et al.*, 1996]. For all other
 805 faults, model slip rates are significantly smaller than
 806 Holocene rates but consistent with geodetic rates within
 807 errors (although typically at the lower edge of the geode-
 808 tic error bar) for the Kunlun [Chen *et al.*, 2000], Central
 809 Altyn Tagh [Bendick *et al.*, 2000; Shen *et al.*, 2001; Wal-
 810 lace *et al.*, 2004], Haiyuan [Lasserre *et al.*, 1999], and
 811 Karakorum [Van Der Woerd *et al.*, 2000; Brown *et al.*,
 812 2002; Chevalier *et al.*, 2004; Wright *et al.*, 2004] faults.

813 We tested whether decreasing the fault friction coef-
 814 ficient would lead to a better match between observed
 815 and model fault slip rates. Since fault friction applies to
 816 all faults in the model, and since India-Eurasia interplate
 817 coupling and convergence are not sufficient to overcome
 818 buoyancy forces in Tibet with a low friction coefficient
 819 on the Himalayan frontal thrust, we treated the India-
 820 Eurasia collision as a displacement-fixed boundary, im-
 821 posing the reference model velocities along that boundary
 822 of the model. All other model parameters and boundary
 823 conditions are kept identical to the best-fit model pre-
 824 sented above. A model with a fault friction coefficient of
 825 0.01 leads to an *RMS* misfit of the predicted surface ve-
 826 locities of 5.2 mm/yr, or 1 mm/yr larger than the best-fit
 827 model described above. Model slip rates are larger than
 828 observed Holocene rates for faults where the reference
 829 model showed a good agreement (Gobi Altay, Baikal,
 830 Tien Shan), but match observed Holocene rates better
 831 for the Jiali and Kunlun faults (Figure 17). They are
 832 however still significantly slower than observed Holocene
 833 rates for the Altyn Tagh, Xianshuihe, and Karakorum
 834 faults.

835 The disagreement between geodetic and Holocene fault
836 slip rates in Asia is highly debated, with estimates that
837 vary by a factor of two to five for the Altyn Tagh and
838 Karakorum faults, for instance [e.g., *Mériaux et al.*, 2004;
839 *Wright et al.*, 2004]. *England and Molnar* [2005] proposed
840 that Holocene slip rates may be biased by an systematic
841 underestimation of terrace riser ages. However, a recent
842 radar interferometry study across the central part
843 of the Altyn Tagh found a slip rate of 18 mm/yr [*Socquet*
844 *et al.*, 2005], in agreement with some Holocene estimates.
845 Although the debate remains open, the dynamic models
846 shown here require rates of 5 to 10 mm/yr on the Al-
847 tyn Tagh fault and are not consistent, overall, with the
848 high slip rate values found by some Quaternary geology
849 studies.

7. Conclusions

850 We have shown that GPS-derived velocities in Asia
851 and, to some extent, Holocene fault slip rates, can be ex-
852 plained by a balance of boundary and buoyancy forces
853 acting on a faulted lithosphere with laterally-varying
854 strength. Model and observed strain rates match well.
855 Both show no resolvable strain rates over most low-
856 elevation regions – which also correspond to areas of
857 high lithospheric strength ($> 10^{23} \text{ Pas}$). Significant in-
858 ternal strain, possibly continuous, is limited to high el-
859 evation areas, mechanically weaker (10^{21} to 10^{23} Pas).
860 Even though block- or plate-like motions provide an ac-
861 curate kinematic description of surface deformation for
862 a large part of Asia [*Calais et al.*, 2006], current GPS
863 strain rates are consistent with the thin-sheet physical
864 framework [*England and Houseman*, 1986], with lateral
865 variations of lithospheric strength controlling the style of
866 deformation.

867 GPS observations are consistent with a model in which
868 subduction boundaries along the eastern and southern
869 borders of Asia (Pacific, Philippines, and Australian
870 plates) play a significant role in the dynamics of defor-
871 mation in Asia [e.g., *Kong and Bird*, 1996]. Our exper-
872 iments show that south China, north China, and Sunda
873 are “pulled” eastward as a result of tensional, ocean-
874 ward stresses rather than “extruded” as a result of India-
875 Eurasia collision. These tensional stresses may result
876 from GPE gradients across the active margins of east-
877 ern and southeastern Asia (as proposed here), or from
878 other subduction processes not accounted for in our mod-
879 els (e.g., slab pull, corner flow). They may also re-
880 sult from viscous coupling between mantle flow and the
881 base of the lithosphere (“basal drag”), a process absent
882 from our models, but that may significantly contribute
883 to the deformation of continents [*Bird*, 1998; *Becker and*
884 *O’Connell*, 2000; *Bokelmann*, 2002].

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Parameters	Weaker rheology		Stronger Rheology	
	Crust	Mantle	Crust	Mantle
Average density at $P = 0$ and $T = 0$ (kg m^{-3})	2816	3330	2816	3330
Temperature at the base of the lithosphere ($^{\circ}\text{C}$)		1200		1300
Coefficient of volumetric thermal expansion (K^{-1})	2.4×10^{-5}	3.94×10^{-5}	2.4×10^{-5}	3.1×10^{-5}
Thermal conductivity ($\text{W m}^{-1} \text{K}^{-1}$)	2.70	3.20	3.50	5.10
Radioactive heat production (W.m^{-3})	7.27×10^{-7}	3.2×10^{-8}	4.6×10^{-7}	negligible

Table 1. Model parameters used to calculate crustal and lithospheric thicknesses (from *Bird* [1998] and *Liu and Bird* [2002b] for the weaker rheology, from *Kong and Bird* [1996] for the stronger rheology tested here).

Parameters	Crust	Lithospheric mantle
A ($\text{Pa s}^{1/n}$)	2.3×10^9	9.5×10^4 *
B (K)	4000	18314
C (K m^{-1})	0	0.0171
σ_s^{max} (MPa)	500	500

* $5.4 \times 10^4 \text{ Pa s}^{1/n}$ for the stronger rheology

Table 2. Non-newtonian dislocation creep law parameters used in the models. These parameters are based on previous deformation modeling in California [*Bird and Kong, 1994*] and Alaska [*Bird, 1996*] for the crust and on experiments on olivine deformation for the mantle [*Kirby, 1983*].

Experiments	bathymetry	Boundary velocities w.r.t. Eurasia				
		Tc (MPa)	To (MPa)	IN, Ar	AUS, Ph, PAC	NAM
1	actual	0	0	-	-	0
2	actual	20	0	0	-	0
3	actual	20	0	REVEL	-	0
4	actual	20	4	REVEL	0	FREE
5	modified	20	4	REVEL	REVEL	REVEL
REF	actual	20	4	REVEL	REVEL	REVEL

Table 3. Summary of parameters used in the experiments reported in section 5 (REF : reference model, section 4.2). For all experiments described above, internal friction coefficient is 0.8 and fault friction coefficient 0.06. Plates: EU, Eurasia; IN, India; Ar, Arabia; AUS, Australia; Ph, Philippines Sea; PAC, Pacific; NAM, North America.