On the plate boundary forces that drive and resist Baja California motion

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ABSTRACT

The driving forces of microplate transport remain one of the major unknowns in plate tectonics. Our hypothesis postulates that the Baja California microplate is transported along the North America–Pacific plate boundary by partial coupling to the Pacific plate and low coupling to the North America plate. To test this idea, we use numerical modeling to examine the interplate coupling on a multiple-earthquake-cycle time scale along the Baja California–Pacific plate boundary and compare the modeled velocity field with the observed geodetic motion of the Baja California microplate. We find that when the strain can localize along a weak structure surrounding microplate (faults), high interplate coupling, produced by frictional tectonic stresses, can reproduce the observed kinematics of the Baja California microplate as seen from geodetic rigid-plate motions. We also find that the northward motion of Baja California can influence the fault slip partitioning of the major faults in the North America–Pacific plate boundary region north of Baja California.

INTRODUCTION

Space geodetic measurements have substantially increased the spatial resolution of the surface velocity field of plate motions over the last decades. Particularly in continental plate boundary regions these data revealed an increasing amount of microplates and rigid blocks. With these subdivisions, the meaning of the word "plate" as an individual mechanical entity for which we can apply concepts such as torque balance (Forsyth and Uyeda, 1975) becomes less clear. In the classical view, the motion of a plate was driven from within by body forces (including ridge push and slab pull), coupling of the plates to the asthenosphere, and frictional sliding along the plate boundaries. However, a common hypothesis is that microplates are externally driven, i.e., that larger neighbor plates determine their motion. The purpose of this paper is to test this "neighbor-driven microplate" hypothesis for the Baja California microplate.

Until the early Miocene, the Farallon plate subducted beneath North America. As the East Pacific Rise approached the trench, subduction of the remnant pieces of the Farallon plate and spreading of the mid-ocean ridge ceased (Lonsdale, 1989, 1991). In the middle Miocene (ca. 12 Ma), extension initiated in a diffuse region along the former volcanic arc ("Protogulf") east of Baja California (Stock and Hodges, 1989). By the end of the Miocene (ca. 6 Ma), the main plate boundary had localized in the Gulf of California, and Baja California was detached from the North America plate (Lonsdale, 1989).

Plattner et al. (2007) used global positioning system (GPS) measurements to show that Baja California is currently moving with respect to North America in approximately the same direction as the Pacific plate, but at a rate that is ~10% slower than the Pacific plate. This result agrees with conclusions from geological studies (Fletcher et al., 2007; Michaud et al., 2004; Nicholson et al., 1994) that Baja California is partially coupled with the Pacific plate. The GPS results also imply that Baja California is only loosely coupled to North America. Here, we address the question of how high Baja California–Pacific coupling stresses need to be, and how low Baja California–North America stresses need to be, to reproduce the regional kinematics. The following geodynamic model allows us to independently constrain these plate boundary forces.

MODEL SETUP

We solve the mechanical equilibrium equations using the finite-element code G-TECTON with (two-dimensional) plane stress spherical shell elements (Govers and Meijer, 2001) and a reference thickness of 100 km. East and west model domain boundaries (Fig. 1) are chosen far from our region of interest. The north-south extent of the model is chosen to encompass the region between the Mendocino and Rivera triple junctions. The model is edge-driven by geodetically constrained velocity boundary conditions (Plattner et al., 2007) relative to North America, with two exceptions: (1) The southern boundary of Baja California is left unconstrained to not impose additional driving or resisting forces, and (2) the northern boundary of the Sierra Nevada microplate is also free to move, because its interaction with adjacent North America is undefined. Plattner et al. (2007) demonstrated that the (GPS) instantaneous Baja California–North America and North America–Pacific velocities are compatible with geologic averages; thus the same edge velocities can be used to drive our mechanical model on both geodetic and geological time scales.

Material properties are homogeneous throughout our model domain for simplicity. On short time scales, the lithosphere between plate boundaries is approximately elastic. On longer time scales, viscous relaxation occurs in the ductile lower crust and upper mantle while the upper crust behaves in a brittle manner (Kohlstedt et al., 1995), i.e., accumulated stresses are thus relaxed by permanent deformation. Over large areas and long time scales this behavior can be approximately represented by a release of stress during a characteristic period (Lambeck, 1988; Stüwe, 2007). In this paper we represent the lithosphere by a single viscoelastic layer. Young's modulus is 75 GPa, and Poisson's ratio is 0.3. We use a reference viscosity of 10^{23} Pa·s, corresponding to a characteristic (Maxwell) relaxation time of 110 k.y.

In our large-scale model, we ignore local details of the fault geometry by only including (micro)plate boundaries along the approximate fault traces (see "Model Sensitivity"). The North America–Pacific plate boundary follows the San Andreas–Gulf of California fault system, and the model fault west of Baja California follows the surface trace of the San Benito–Tosco Abreojos fault system, which is the former trench. The Sierra Nevada–North America boundary is only well defined south of 39.5°N, along the Walker Lane–Eastern California shear zone. North of this latitude the location of this boundary is unclear (Unruh et al., 2003; Wesnousky, 2005). Hence, our model Sierra Nevada–North America boundary extends only as far north as 39.5°N (see "Model Sensitivity"). Shear along the San Jacinto and Elsinore faults is simulated by a single fault.

All model faults are vertical (the implication of a vertical Pacific–Baja California plate boundary is explained later), and strike slip can occur in response to shear stress on model faults (Melosh and Williams, 1989). Seafloor spreading at Gulf of California ridges south of 27°N is modeled by allowing both strike-slip and normal relative motion. Fault intersections are represented by triple overlapping nodes. Most model faults are frictionless, besides the San Benito–Tosco Abreojos fault. We investigate mechanical coupling of the Pacific plate with Baja California by varying the dynamic friction along the San Benito–Tosco Abreojos fault. This shear stress magnitude is the key parameter of this work.

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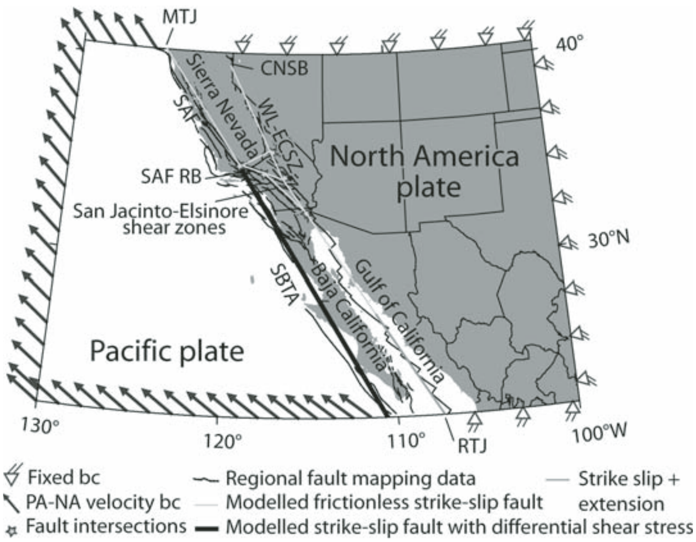

Figure 1. Model domain with velocity boundary conditions for fixed North America (velocity vectors length not to scale). The figure shows the major tectonic blocks used in the model. The Baja California microplate is coupled with the Pacific plate along their common plate boundary. SAF—San Andreas fault; SAF RB—San Andreas fault restraining bend; CNSB—Central Nevada Seismic Belt; WL-ECSZ—Walker Lane–Eastern California shear zone; SBTA—San Benito–Tosco Abreojos fault system; MTJ—Mendocino triple junction; RTJ—Rivera triple junction; bc—boundary condition; PA-NA—Pacific–North America. Regional faults from (Jennings, 1994) and INEGI (http://galileo.inegi.gob.mx/website/mexico/).

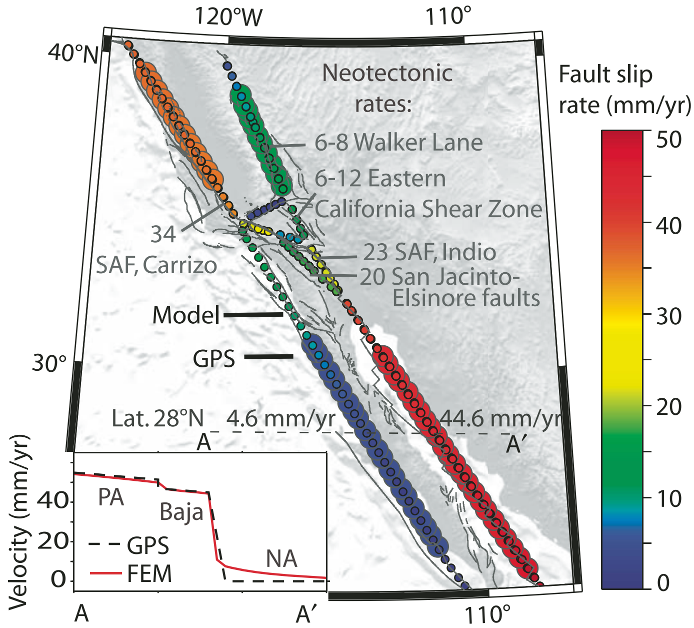

Figure 2. Regional fault kinematics for an optimal shear stress applied along the San Benito–Tosco Abreojos fault scaled to fit the model velocity (FEM) to the observed geodetic (GPS) rigid-plate motion along a profile at latitude 28°N (inset). The model fault velocity (small dots) around rigid Baja California fits the corresponding observed geodetic rigid-plate relative motions (large dots). The motion of the Sierra Nevada microplate induced by the collision of Baja California is also in good agreement with the geodetic rates. Fault slip rates around the San Andreas fault (SAF) restraining bend agree with neotectonic rates for the corresponding fault system (summarized in Becker et al., 2005).

MODELING RESULTS

We first concentrate on the short-term response of the model, for comparison with observations on geodetic time scales (tens of years). This model is essentially elastic. Next, we look at geological time scales (beyond 100 k.y.), to examine the consequences of permanent deformation on plate coupling.

Plate Coupling on Geodetic Time Scales

To test which fault coupling causes surface displacements that agree with GPS observations, we increase the dynamic friction along the San Benito–Tosco Abreojos fault from zero (i.e., free slip) until the model Baja California microplate becomes fully locked to the Pacific plate. Along a profile at latitude 28°N (Fig. 2) we look for the frictional shear stress value that optimizes the fit between the model kinematics and the geodetic observations (Fig. 2). For a frictionless San Benito–Tosco Abreojos fault (model not shown here), the partitioning of Pacific and North America plate motion among the faults bounding Baja California is controlled solely by the fault geometry. Approximately half of the total North America–Pacific rate is accommodated on the San Benito–Tosco Abreojos fault system, and half along the Gulf of California. When we increase the frictional shear stress along the San Benito–Tosco Abreojos fault system, fault slip increasingly concentrates on the Gulf of California until the full North America–Pacific relative motion is accommodated east of Baja California (fully locked Baja California).

The model that gives the best fit with the kinematic observations along 28°N is obtained when the friction is 90% of the locking shear stress. For this coupling, the differential motion along the (micro)plate-bounding faults of our numerical model (small dots in Fig. 2) is in good agreement with the relative velocities computed from the corresponding Euler poles (Plattner et al., 2007; Psencik et al., 2006) (large dots in Fig. 2). A comparison of single model velocity vectors with observed GPS velocities within the rigid Baja California and Sierra Nevada microplates (Plattner et al., 2007; Psencik et al., 2006) shows agreement within the measurement uncertainty (Fig. 3).

When the deformation can localize along a weak structure surrounding the microplate (in our case the Gulf of California ridge-fault system), Baja California keeps its geodetic rigidity while being transferred by partial coupling to the Pacific plate. However, a continuum of deformation occurs in regions bounding the model faults (even for a frictionless San Benito–Tosco Abreojos fault). This (elastic) strain increases with time (or fault slip) because of the velocity boundary conditions that drive the model. The stresses associated with these strains represent a continuum of resistance to microplate motion due to (partial) misalignment of the faults with respect to small circles of the plate rotation poles.

To balance the increasing elastic resistance while keeping the relative plate motions constant, the frictional coupling in the model needs to increase with time. For the best-fit model, we need to increase the frictional shear stress along the San Benito–Tosco Abreojos fault by 91 Pa/yr.

Deformations of the elastic model presented above can only be realistic on a short time scale (approximately one earthquake cycle). On longer time scales it is unlikely that the lithosphere can accumulate infinite stress at the plate boundaries. Therefore, in the next section, we allow for stress relaxation.

Plate Coupling on Geologic Time Scales

On time scales that are comparable to, or longer than, the characteristic relaxation time, stress flow (partly) reduces the continuum of resistance (Fig. DR4 in the GSA Data Repository[1]). To maintain the fit with observations, successive coupling stress increments must therefore decrease with

[1]GSA Data Repository item 2009088, stress relaxation in the viscoelastic model, is available online at www.geosociety.org/pubs/ft2009.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.

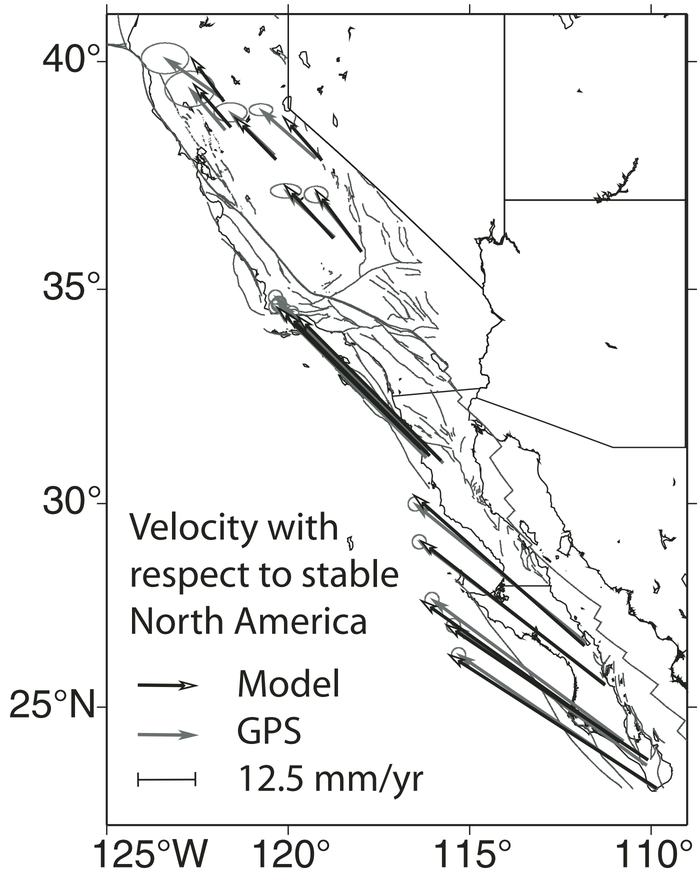

Figure 3. Model velocity vectors and GPS velocity vectors from sites located within the rigid Baja California and Sierra Nevada microplates and used for the Euler vector calculations (Plattner et al., 2007; Psencik et al., 2006).

time. As a matter of fact, the increment of shear stress decays exponentially at a rate that is controlled by the Maxwell time and by the total applied shear stress along the San Benito–Tosco Abreojos fault. The eventual steady state represents dynamic equilibrium between the stress increase by loading and decrease by viscous relaxation during a given period. The steady state is generally reached after a loading period of approximately ten Maxwell times. Because of the exponential decay of the applied incremental shear stress, the magnitude of the steady-state shear stress along the San Benito–Tosco Abreojos fault is equivalent to the initial (elastic) increment multiplied by the Maxwell relaxation time. For the selected Maxwell time of 110 k.y., the steady-state shear stress applied to the San Benito–Tosco Abreojos fault is 10 MPa. For a longer Maxwell time, a higher steady-state shear stress is required to activate Baja California motion.

MODEL SENSITIVITY

Where the geometry and location of the most important fault zones in the study area (Fig. 1) are incompletely known, our choice was motivated by better fit to the GPS observations. Our coupling stress is partly affected by these uncertainties. The coupling stress scales inversely with the length of the Sierra Nevada–North America plate boundary as the fault length affects the friction between the Sierra Nevada microplate and North America (10% stress variation for length changes within the Central Nevada Seismic Belt). Westward shift of the northern San Benito–Tosco Abreojos fault and more northerly connection to the San Andreas fault allow Baja California to pass west of the restraining bend, shifting motion to the San Andreas fault and the Sierra Nevada microplate, and lower the coupling stresses (4 MPa for a model in which the northern San Benito–Tosco Abreojos fault continues in the western California borderland and connects along the San Gregorio fault to the San Andreas fault). The coupling stress estimate is insensitive to representing the Gulf of California by a sequence of small ridges or basins connected by long transform faults. However, varying the length of the segment that permits an extensional component of slip perpendicular to the Gulf of California boundary can affect the stress by up to 30% (strongly affecting the Baja California velocity azimuth). We did not test how much resisting forces can be lowered by including several smaller fault traces around the restraining bend to simulate continuously deforming areas and the effect of smaller blocks such as the Western Transverse Ranges taking up deformation by block rotation. However, we found that not including the San Jacinto or Elsinore fault causes Baja California to rotate clockwise more than observed by GPS velocity azimuths.

In the above models, we assumed that the Gulf of California fault is weak. When we increase coupling across this fault, stresses along the San Benito–Tosco Abreojos fault need to be increased by the same amount to maintain the fit to the geodetic Baja California velocities. Coupling stresses in the Gulf of California beyond 5–10 MPa reduce fault slip to the extent that internal deformation starts occurring within Baja California. Detecting such (permanent) internal deformation using a GPS is complicated by earthquake cycle signatures from the plate boundaries and the faults located within the GPS network in northern and southern Baja California. As coupling stresses probably decreased over geological time with the progressive localization of strain along Gulf of California ridges and transforms, we would expect that the geodetic strain rate within Baja California of 10^{-16} s^{-1} (Plattner et al., 2007) is lower than average strain rates within Baja California estimated from geology.

The required shear stress to reproduce Baja California motion strongly depends on the length of the coupled segments along the San Benito–Tosco Abreojos fault. Our analysis of geodetic motions only gives the average coupling stress. Reducing this length (while still preserving the velocity field) increases the necessary coupling stress at these segments, and in consequence would introduce local deformation and rotation within Baja California. From geological history it is possible to argue for variations in coupling between fracture zones (Michaud et al., 2006).

ANALYSIS

In our best-fit model, we require a shear stress of 10 MPa on the San Benito–Tosco Abreojos fault on geological time scales. This stress magnitude is typical for tectonic stress in the lithosphere (Gardi et al., 2003; Iaffaldano and Bunge, 2008), indicating that Baja California being dragged by the Pacific plate is a plausible mechanism. The frictionless-fault assumption for all plate boundaries besides Pacific–Baja California and the block model geometry yield a lower limit for the lithospheric coupling.

The motion of Baja California induced by coupling with the Pacific plate activates faults surrounding the Sierra Nevada microplate. Although the Sierra Nevada microplate is not fully decoupled from North America, the induced fault slip rates along the southern Sierra Nevada plate boundaries are in good agreement with the observed geodetic rigid-plate motion of the Sierra Nevada microplate with respect to the Pacific and North America plates (dots in Fig. 2), and with the microplates' velocity field (vectors in Fig. 3) (Psencik et al., 2006).

The northward motion of Baja California influences the dynamic behavior of the restraining bend fault system (Fig. 2). Here, a comparison of the model velocity field with GPS observations is not possible since interseismic strain accumulation affects the geodetic data. Regional neotectonic fault slip rates that represent deformation averaged over at least ten earthquake cycles (Becker et al., 2005) indicate a reasonable fit for the main fault traces (Fig. 2).

Our model predicts significant vertical strain along the San Andreas restraining bend, where contraction leads to crustal thickening, and along the Gulf of California, where we predict thinning when fault-perpendicular motion is not permitted. The modeled regional pattern of vertical motions is in good agreement with geologic observations. However, rates of crustal thickness change, and equivalent uplift or subsidence are significantly smaller than observed. This can be expected since we use a linear rheology that does not allow for localization of the deformation and we do not include effects such as weakening from a pre-existing volcanic arc in the Gulf of California (Fletcher et al., 2007).

DISCUSSION AND CONCLUSIONS

The northward migration of Baja California due to coupling with the Pacific plate is hindered in the north by the presence of the Sierra Nevada block and its coupling with North America. For this reason, development and formation of the eastern boundary of the Sierra Nevada block play a fundamental role in the forces necessary to drive Baja California. An alternative for resisting forces induced by the Sierra Nevada block is the presence of a slab window beneath the Baja California microplate lowering the coupling to the Pacific plate. However, such a model is only possible with three-dimensional geometry.

A mechanism to increase the coupling along the former plate boundary between the North America and Pacific plates was suggested by Nicholson et al. (1994). Cessation of spreading between fragments of the former Farallon plate and the Pacific plate left slivers of young oceanic lithosphere stalled beneath Baja California. Remnants of these slab fragments beneath Baja California extending all the way out into the Gulf of California are interpreted from seismic tomographic images (Zhang et al., 2007). If indeed coupling is due to these stalled microplate fragments, the interface is probably inclined. Because, in our model, we constrained the coupling stress for a vertical interface, the real area of the interface would be larger and thus would require lower shear stress to generate equivalent frictional forces.

The reproduced kinematics of Baja California and the North America–Pacific plate boundary region, when Baja California is driven by mechanical coupling between Baja California and the Pacific plate through tectonically reasonable stresses, suggests that this process is a plausible mechanism for driving the transport of the microplate. Partial coupling and resisting forces to the northward migration of Baja California can be an explanation for the current motion of the peninsula in the same direction but at a smaller rate than the Pacific plate.

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