

Mantle Control of Plate Boundary Deformation

Tim Melbourne¹ Don Helmberger²

Abstract. The seismic wavefield propagating along the recently instrumented Pacific-North American plate boundary (California) displays remarkable variation, with regional shear waves arriving at coastal stations up to 20 seconds earlier than equidistant stations in eastern California. Broadband modeling of this data reveals that coastal paths sample fast upper mantle typical of Miocene-aged ocean plate (> 50 Km thickness). Inland paths sample slower uppermost mantle, with the seismic lithosphere, or lid, measuring less than 5 Km thick, characteristic of the Basin and Range extensional province. The boundary in the uppermost mantle between these provinces is sharp, expressing the juxtaposition of the stronger Pacific plate with weaker continental North America. The lid step coincides with regionally maximum dextral strain rates measured with GPS, suggesting the uppermost mantle modulates long term, regional-scale continental margin deformation and evolution.

Introduction

What controls crustal deformation along active continental margins? In the non-plate dynamics of distributed continental strain, one school of thought would attribute active control to local buoyancy forces intrinsic to the lithosphere [Jones *et al.*, 1996]; [England and Molnar, 1997], while another would argue the sub-crustal mantle generates plate boundary forces which control not only margin deformation but strain interior to continents as well [Wernicke and Snow, 1998]. Field evidence would seem to support both hypotheses: regional minimization of crustal potential energy has been successful in predicting some localized regimes of distributed continental extension, but palinspastic reconstructions of the Basin and Range province, the archetype of distributed continental extension, show clear changes in extension direction which closely track simultaneous shifts in the relative motion of the Pacific plate [Atwater and Stock, 1998]. Delineating the importance of these forces requires mapping margin lithospheric thickness, a generally difficult task due to the low velocity zone which traps lower-lithosphere seismic phases. However, new instrumentation of the Pacific-North American plate boundary with dense broadband seismometers is now allowing tracking of previously unobservable diffracted phases along the continental margin which provide precise constraints on lateral variation in the lithospheric lid. Here the term lid refers to the seismic lithosphere, or that part of the sub-crustal upper

mantle whose fast velocities delineate the elastic lithosphere [Anderson, 1989].

Unlike teleseismic phases, regional S^2 is particularly sensitive to lid structure. Propagating along the North American margin between the southern Baja Peninsula and Northern California, S^2 travels mostly horizontally and spends the majority of its travel path sampling the uppermost mantle. Both travel time anomalies and waveforms here are strongly azimuthally dependent. The phase S^2 , timed relative to S to sidestep event location and onset errors, is consistently early along the most western raypaths, with coastal stations advanced relative to the shear wave model Tectonic North America (TNA), appropriate for the Basin and Range [Grand and Helmberger, 1984] (Figure 1). In contrast, nearby stations in eastern California are well predicted by this model.

Synthetics were generated with frequency - wavenumber (FK) Green's functions to accurately model tunneled energy propagating in the sub-lid low velocity zone. Synthetic sources were placed at 8 Km depth, consistent with shallow transform events, and Green's functions were linearly combined according to Harvard CMT moment tensors [Dziewonski *et al.*, 1997]. The East Pacific Rise-specific attenuation model of [Ding and Grand, 1993] improved the relative $S - S^2$ amplitude fits over synthetics computed without an attenuation model. The data and synthetics are aligned on S , which eliminates the need for precise origin times and epicenters [Grand and Helmberger, 1984]. We employed a grid search of different lid thicknesses, holding the path-averaged crustal thickness constant.

S^2 is triplicated by the 410 Km discontinuity, forming twin-peaked waveforms beyond 31° . At distances closer than 36° , the first arriving branch of the triplication bottoms in the shallow mantle, and lithospheric lid thickness strongly influences $S^2 - S$ travel time and S^2 waveform shape via interference of the two triplication branches [Helmberger *et al.*, 1985]. The velocity increase associated with the lithospheric lid ($V_s = 4.55$ Km/S) represents a 5% increase over non-lithospheric upper mantle velocities ($V_s = 4.3$ Km/s). Large $S^2 - S$ travel time anomalies are generated in synthetics with only modest increases in lid thickness, providing a sensitive parameterization. For instance, the shallow mantle branch of S^2 advances over 20 seconds by increasing the velocity of the upper 55 Km of the mantle by 5% from TNA, as displayed in Figure 1.

S and S^2 waveform fits play an equal role with travel times in constraining lithospheric lateral variation. The primary waveform diagnostic is S^2 branch interference, which is tightly controlled by varying lid structure (Figure 2). For instance the double-peaked shapes at MHC (31.5°) and ORV (33.2°) are caused by relative phase advancement which can be well modeled with lid thicknesses of 48 Km and 8 Km, respectively. For each path traversing coastal margins, the synthetic arrival does not split until the seismic lithosphere reaches thicknesses in excess of 40 Km. For easterly paths,

¹Department of Geological Sciences, Central Washington University

²Seismological Laboratory, California Institute of Technology

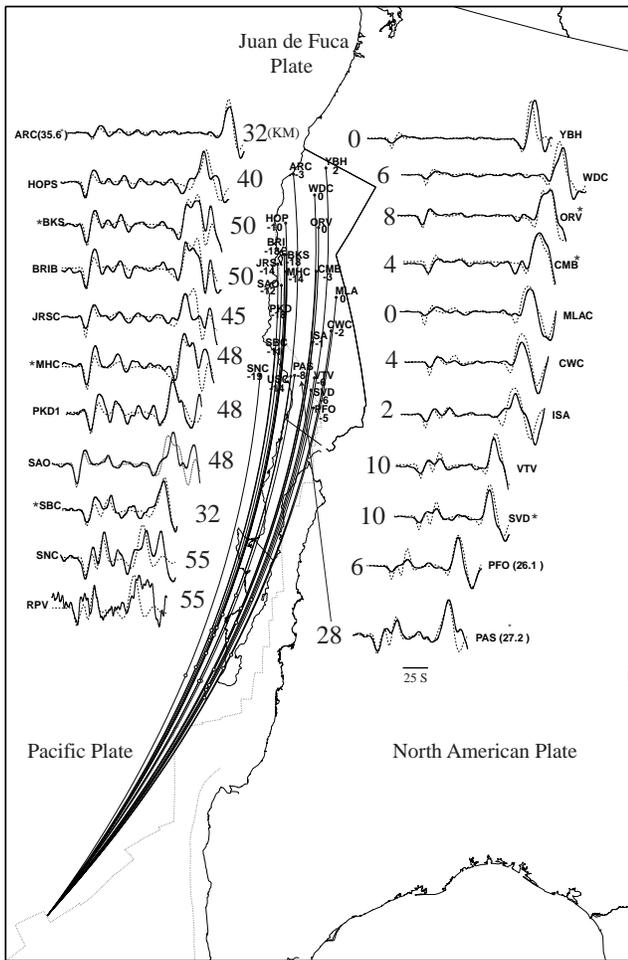


Figure 1. Observed (solid) and calculated (dashed) shear body waveforms recorded along Pacific-North American plate boundary, with corresponding lid thickness printed beside each trace. Note that S^2 arrives over 10 seconds earlier at MHC compared to equidistant CMB (31.5°). Their paths are separated by only 50 Km at the S^2 midpoint bounce, and 100 Km near Pasadena, whose station (PAS) has an average delay and corresponding lid thickness. Westernmost raypaths uniformly require substantially thicker lithospheric lid structure, averaging 50 Km, while eastern paths require thin or no lithospheric lid structure, consistent with known Basin and Range upper mantle shear structure. $S^2 - S$ travel time anomalies relative to TNA Grand and HelMBERGER [1984] in seconds are shown below station name. FK synthetics are computed for 1D average lithospheric lid thicknesses indicated adjacent to waveform traces. S^2 midpoints (white circles) all lie beneath the topographic expression of peninsular Baja California. Stations with asterisks are detailed in Figure 2.

the singular arrival in the data mandates lids less than 10 Km.

Neither the free surface reflection nor the source-side leg of S^2 raypaths can be responsible for the large travel time and waveform anomalies observed in the data, for several reasons. First, ray tracing indicates that all S^2 phases shown here bounce beneath the topographic expression of the Baja California Peninsula; none bounce beneath the Pacific ocean floor. Regardless, sub-oceanic bounce points could only provide 2 seconds two-way travel time shift, only 10% of the maximum observed anomaly. Second, 20 second dominant frequencies indicate that wavefront healing effects are

sufficiently large that differential times can not be accrued along the source leg, even for extreme velocity heterogeneity [Frankel and Clayton, 1986]. Furthermore, along this leg the raypaths traverse primarily beneath Miocene ocean floor, known to show little lateral variation except directly beneath the East Pacific Rise spreading center [Melbourne and HelMBERGER, 2000]; [Forsyth *et al.*, 1998]. Third, absolute travel times of S show none of the travel time anomalies observed with S^2 , indicating that the anomalous velocity structure must be distributed along shallower paths not sampled by S . Together, these features confine the anomaly to be distributed along the North American margin between northern California and the Baja California Peninsula. Finally, this margin is known to be complex from a host of independent measurements, including surface wave dispersion measurements across the California margin which show uppermost mantle structure equivalent to the average of lid thicknesses reported here [Polet and Kanamori, 1997], receiver function analysis [Zhu and Kanamori, 2000], and absolute travel times in teleseismic tomography [Humphreys and Dueker, 1994].

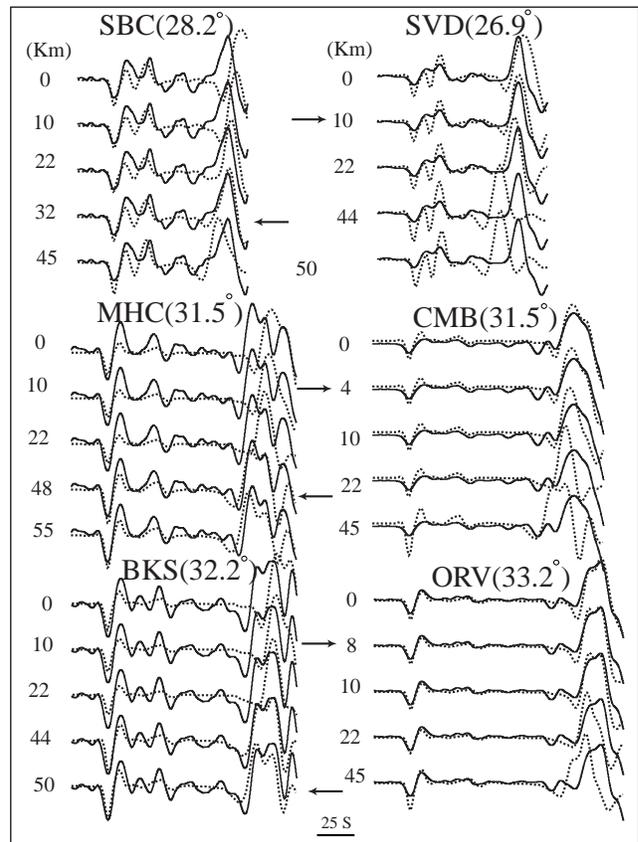


Figure 2. Influence of Lithospheric lid thickness on $S^2 - S$ travel times and S^2 waveforms. Varying lid thickness between 0 and 50 Km produces over 20 seconds change in $S^2 - S$ travel time, and strongly alters S^2 pulse shape. Each shear wave seismogram is repeated 5 times, and overlaid with synthetics with a single 1D model in which the thickness of the high velocity lid ($V_s=4.55$ Km/s versus $V_s=4.4$ Km/s non-lid) varies according to thickness printed to left of data trace. Lid thickness providing minimal synthetic-data misfit is identified with arrow.

Implications for Plate Boundary Tectonics

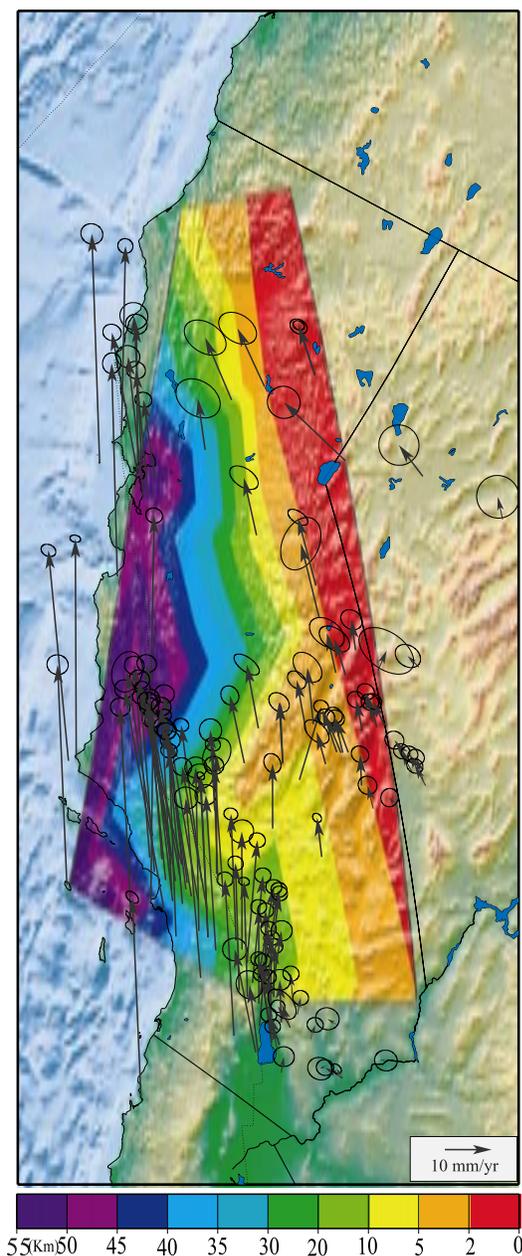


Figure 3. Contoured seismic lithosphere thicknesses beneath North American continental margin derived from broadband S^2 waveforms and $S^2 - S$ travel times. Lithospheric lid ($V_s = 4.55$ Km/S versus 4.3 Km/S non-lid, TNA upper mantle) thicknesses range systematically from 55 km along margin coastal paths (Peninsular Baja-Western California), typical of Miocene-aged oceanic lithosphere and similar to that observed beneath eastern Pacific, to effectively 0 Km along Eastern California-Eastern peninsular Baja paths, typical of Basin and Range upper mantle. Intraplate dextral strain, indicated by geodetic measurements conducted across thick lithospheric lid regions, is substantially lower than interplate dextral strain, indicated by measurements taken across regions of high lid thickness gradients. Continental margin field vectors, derived from SOPAC archives Bennett *et al.* [1999], Eastern California Shear Zone Miller *et al.* [2000], and Sierran block measurements Dixon *et al.* [2000] are relative to related definitions of stable North America (co-occupied GPS vectors found to differ less than 15% among the three studies).

The presence of a step function in uppermost mantle rigidity along the ocean-continent boundary should have a profound, if not controlling, influence on the long term evolution of the plate margin. Geodetic measurements indicate that current margin deformation occurs within a broad zone of strike-slip accommodation penetrating at least 400 Km into the continental interior, but the majority of shear strain occurs in close proximity to the surface expression of the continental margin (Figure 3; [Bennett *et al.*, 1999]; [Lisowski *et al.*, 1991]). The correlation between lithospheric thicknesses imaged here and GPS-determined dextral strain rates suggest the weaker North American continental crust accommodates via deformation the relative motion of the stronger Pacific lithosphere. Dextral strain calculated from velocity profiles across uniformly thick lithospheric lid regions is substantially lower than dextral strain calculated from profiles across regions of high lid thickness gradients. For instance, along an E-W transect across the lid step, displacement rates relative to North America drop from 40 mm/yr at coastal stations to 5 mm/year east of the Sierran block and Eastern California Shear Zone [Miller *et al.*, 2000]; [Dixon *et al.*, 2000]; [Bennett *et al.*, 1999]. Vectors show that California coastal region, underlain by the thickest lithosphere, largely moves as an undeforming rigid structural block.

Observed motion of crustal massifs such as the Sierra-Great Valley block, known to displace without internal deformation [Dixon *et al.*, 2000], will reflect the sum of forces exerted along them by a variety of processes. If basal tractions are proportional to lithospheric rigidity, as recent studies suggest in Southern California and elsewhere [Bourne *et al.*, 1998], then surface observations of block-like motion may belie lateral variation within the sub-crustal mantle. In this case, major block boundaries such as the San Andreas and its sympathetic faults then accommodate differential crustal motion driven by laterally varying mantle rigidity, rather than specifically delineating the plate boundary at depth.

Strong oceanic lithospheric mantle can also buttress the proximal, weaker continental lower crust and by doing so modulate interior continental tectonics. Known changes in Basin and Range extension directions at 8 and 13 MA [Wernicke and Snow, 1998] track simultaneous changes in relative Pacific-North American plate motion [Atwater and Stock, 1998], which likely reflects such modulation. Continental material, extending due to intrinsic buoyancy forces [Jones *et al.*, 1996], is nonetheless spatially confined by stronger Pacific plate lithosphere. Present day extension within the Basin and Range is NW-directed, largely towards the Cascadia arc and away from the Pacific-North American transform boundary [Bennett *et al.*, 1999], [Thatcher *et al.*, 1999]. If the Pacific lithosphere constitutes a translating, impenetrable barrier to lower crustal flow which 'deflects' extending continental material northward, then mass balance requires that the mantle component of the young, subducting Gorda and Juan de Fuca plates must be substantially more penetrable than the impenetrable rheological step imaged here across the lower California transform system.

Acknowledgments. This work was supported under National Science Foundation grants EAR-9973191 to Melbourne and

EAR-97-2508 to Helmberger. Caltech Seismological Laboratory Contribution #8762

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Tim Melbourne Department of Geological Sciences Central Washington University Ellensburg, WA 98926 (email: tim@geology.cwu.edu)

Don Helmberger, Seismological Laboratory 252-21, California Institute of Technology CA 91125 (email: helm@gps.caltech.edu)

(Received March 9, 2001; accepted July 11, 2001.)