Plate Boundary Zones: Concept and Approaches

Seth Stein and Giovanni F. Sella

Department of Geological Sciences, Northwestern University, Evanston IL 60208

An important evolution in ideas about plate tectonics is the recognition that boundaries between plates are often broad zones, rather than the idealized narrow boundaries assumed in the rigid plate model. Initial evidence for this concept came from the distribution of seismicity, active faulting, and topography that showed a broad zone of deformation along a number of plate boundaries. In recent years space geodetic studies have confirmed the idea of broad boundary zones by showing that on many boundaries a significant fraction of the plate motion occurs at considerable distances from the nominal boundary. In these cases, instead of viewing the zone as a narrow boundary bordered by "intraplate" earthquakes and deformation, it is regarded as a broad boundary zone within which the earthquakes and deformation reflect the distribution of motion between the interiors of two major plates. In this view, a boundary zone occurs where motion is distributed over a zone significantly broader than the region of elastic deformation associated with the earthquake cycle along the nominal plate boundary. Conversely, we view the plate interior as the region within which site velocities are well described by rotations about a single Euler pole. Space geodesy, combined with seismology, topography, and structural studies, is giving increasingly better views of how motion within plate boundary zones varies in space and time. In different areas, studies of plate boundaries find a range of kinematic behavior from diffuse deformation zones to discrete microplates. The kinematic data in turn provide valuable information about plate boundary mechanics. For example, they give insight into the complex issues of how motion at convergent plate boundary zones is divided into seismic and aseismic motion at the trench, the primary plate interface, and deformation of the overriding plate. They similarly give insight into the mechanics of uplift and mountain building. These studies have implications for society because about 40% of Earth's population lives within plate boundary zones, and are thus vulnerable to geologic hazards not just at the nominal boundary, but also over the broad boundary zone.

INTRODUCTION

The central concept of the theory of plate tectonics, the primary intellectual underpinning of the modern earth sciences, is that large regions of the earth's surface move coherently, such that most deformation, seismicity, volcanism, and extreme topography occur at their boundaries. In his classic paper introducing transform faults, *Wilson* [1965] suggested that "plates between mobile belts are not readily deformed except at their edges." Although this model of rigid plates with narrow boundaries provided much of the power of plate kinematics, allowing the derivation of models by Euler vector summation for plate circuits, it was recognized to oversimplify continental complexities. *Morgan* [1968] noted that "Such features as the African rift system,

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Cameroon trend, and Nevada-Utah earthquake belt are most likely the type of distortion denied in the rigidity hypothesis. Nonetheless, it is of interest to see how far this simplifying concept of rigidity can be applied."

Using the rigid plate assumption, models for global relative plate motion were derived by combining data along plate boundaries. Rates of plate motion inferred from magnetic anomalies at midocean ridges, together with directions of motion inferred from the azimuths of transform faults and earthquake slip vectors at plate boundaries, were inverted to yield models of geologically instantaneous (averaged over the past few million years) motion between plates [*Morgan*, 1968; *Le Pichon*, 1968; *Chase*, 1972, 1978; *Minster et al.*, 1974; *Minster and Jordan*, 1978; *DeMets et al.*, 1990, 1994]. Such models consist of a set of angular velocity vectors (Euler vectors) specifying the motion of each plate relative to one arbitrarily fixed plate. This approach assumes that plate interiors deform at a low (ideally zero) rate compared to motion between them, which occurs only at narrow boundaries. Hence Euler vectors derived from data from different boundaries can be combined by vector addition.

Successive models have been developed using more and better data, and verified using independent data provided by space-based geodetic systems [*Smith and Turcotte*, 1993]. These include Very Long Baseline radio Interferometry (VLBI) [*Clark et al.*, 1987], Satellite Laser Ranging (SLR) [*Smith et al.*, 1990], the Global Positioning System (GPS) [*Dixon*, 1991; *Segall and Davis*, 1997], and the Doppler Orbitography and Radiopositioning Integrated Satellite System (DORIS) [*Cazenave et al.*, 1992]. These techniques combine precise satellite-based timing, ranging, and orbit estimation to measure the positions of geodetic monuments to centimeter and better accuracy, such that measurements over time yield precise relative velocities.

By using sites in the interior of different plates, space geodesy avoids the assumption of plate rigidity and in fact permits the rigidity to be quantified. The different space-based systems yield comparable velocity data, which can be combined. Because of the relatively lower cost and portability, most of the data presently used are from GPS. (Another space geodetic technique, Synthetic Aperture Radar interferometry (InSAR) from satellites [*Bürgmann et al.*, 2000], shows relative motion within an image tens to a hundred kilometers across, but not absolute positions on a plate-wide or global scale. InSAR is thus used for earthquake or fault studies, but not for larger-scale applications.)

Velocities estimated from space geodesy, which span only a few years, are generally similar to those predicted by global plate motion models that span several million years [*Robbins et al.*, 1993; *Argus and Heflin*, 1995; *Dixon and Mao*, 1997; *Larson et al.*, 1997; *Sella et al.*, 2002]. This agreement is consistent with the prediction that episodic motion at plate boundaries, as reflected in occasional large earthquakes, gives rise to steady motion in plate interiors due to damping by the underlying viscous asthenosphere [*Elsasser*, 1969].

Despite the power of the rigid plate model, it was recognized that boundaries between plates are often broad, rather than the idealized narrow boundaries assumed in the rigid plate model. The initial evidence for this concept comes from the distribution of seismicity and the topography, which often imply a broad zone of deformation between plate interiors. This effect is especially evident within continents, such as the India-Eurasia collision zone in the Himalayas, the Pacific-North America boundary zone in the Western U.S., or the Nubia-Somalia boundary at the East African rift. Broad boundaries also occur at active ocean-continent margins [*Stevens et al.*, this volume; *Tandon et al.*, this volume] and in oceanic lithosphere, as in the central Indian Ocean or along some major transforms [*McGuire et al.*, this volume]. Plate boundary zones, as indicated by earthquakes, volcanism, and other deformation, appear to cover about 15% of the Earth's surface (Figure 1) [*Gordon and Stein*, 1992; *Stein*, 1993; *Gordon*, 1998; 2001].

Plate boundary zones, although recognized prior to the advent of space geodesy, were difficult to study because plate motion models were derived using only data from plate boundaries and thus predict only the integrated motion across them. In recent years, however, space geodetic studies like those discussed in this volume have confirmed the idea of boundary zones, are resolving the motion within them, and are providing key data for investigations of their mechanics. A common theme emerging is that often (but not always) a nominal plate boundary, such as the San Andreas fault, accommodates the largest single portion of the relative plate motion, and is the locus of the largest earthquakes. However, a significant fraction of the motion and hence seismicity occurs away from the nominal boundary (Figure 2). Thus instead of viewing the region as a single plate boundary fault surrounded by "intraplate" earthquakes and deformation, it is regarded as a plate boundary zone within which earthquakes and deformation reflect the distribution of motion between the interiors of two major plates. We thus regard a plate boundary as a broad zone when its width significantly exceeds that of the region of elastic deformation associated with the earthquake cycle along the nominal plate boundary.

Space geodesy, especially when combined with seismological, topographic, and structural studies, is giving increasingly better views of how motion within plate bound-

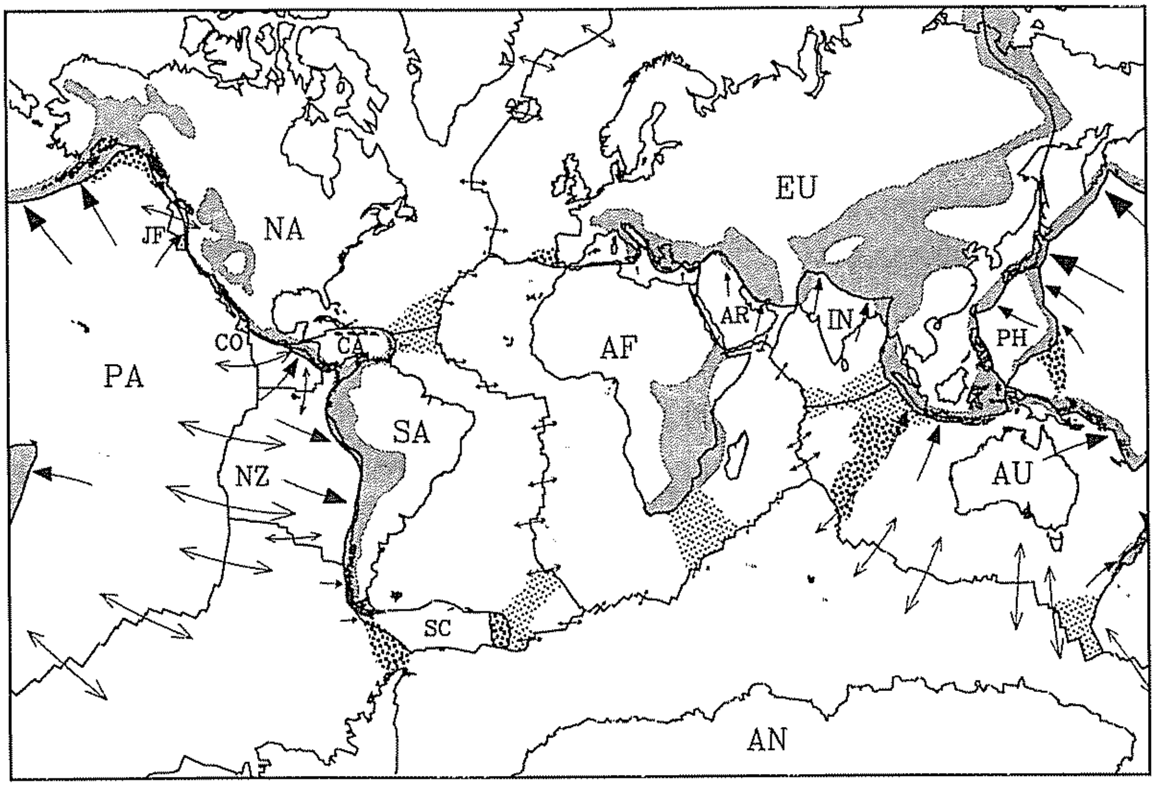

Figure 1. Plate boundary zones and relative plate motions for the NUVEL-1 global plate motion model Arrow lengths are proportional to the displacement if plates maintain their present relative velocity for 25 Myr Divergence across mid-ocean ridges is shown by diverging arrows Convergence is shown by single arrows on the underthrust plate Plate boundaries are shown as broad zones implied by seismicity, topography, or other evidence of faulting Fine stipple shows mainly subaerial regions where the deformation has been inferred from seismicity, topography, other evidence of faulting, or some combination of these Medium stipple shows mainly submarine regions where the nonclosure of plate circuits indicates measurable deformation; in most cases these zones are also marked by earthquakes Coarse stipple shows mainly submarine regions where the deformation is inferred mostly from seismicity The geometry of these zones, and in some cases their existence, is under investigation [*Gordon and Stein*, 1992]

ary zones varies in space and time Possible spatial variations include a single fault system taking up most of the motion [e.g *Prescott et al.*, 1981], a smooth distribution of motion [e.g *England and Jackson*, 1989], or motion taken up by a few relatively large microplates or blocks [e.g *Beck*, 1980; *Engeln and Stein*, 1984; *Acton et al.*, 1991; *Thatcher*, 1995; *Stein et al.*, this volume] Each of these possibilities occurs, sometimes within the same boundary zone, as discussed shortly The distribution of the motion in time is of particular interest because steady motion between plate interiors gives rise to episodic motion at plate boundaries, as reflected in occasional large earthquakes or, in some cases, steady creep The relation between plate motions and earthquakes is complicated and poorly understood and hence forms a prime target of study

Our purpose here is to provide an overview of some of the approaches, issues, and results emerging from plate boundary zone studies This paper is not intended as a comprehensive review simply because of the volume of such studies.

TECHNIQUES FOR STUDYING PLATE BOUNDARY ZONES

Studies of plate boundary zones seek to describe the kinematics, or motions, within plate boundary zones as functions of space and time The kinematic results are then used to draw inferences about the mechanics of boundary zones and of large-scale plate motions These studies combine various types of data, summarized in Figure 3 The data have different strengths and limitations, sample deformation different-

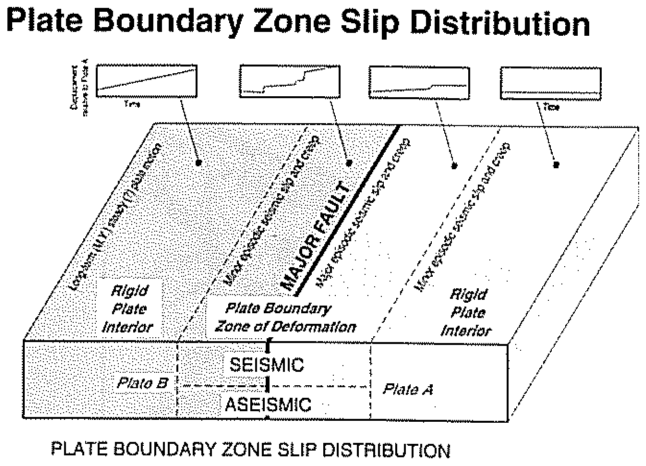

Figure 2. Schematic illustration of the distribution of motion in space and time for a strike-slip boundary zone between two major plates [*Stein*, 1993]

ly over different time scales, and thus provide complementary information.

Before discussing results, we briefly consider the different techniques and note some of their limitations imposed by uncertainties in the data, possible systematic errors, and intrinsic features of the technique. These issues are coming under increasing scrutiny now that diverse data types can be compared and combined to draw tectonic inferences. For example, it is important to decide when present-day space geodetic motions differ significantly from those predicted by models that average over the past few million years, given that the uncertainties assigned to Euler vectors in the models appear to be systematically overestimated [*Minster et al.*, 1974; *Minster and Jordan*, 1978; *DeMets et al.*, 1990] and that assessing the precision of GPS velocities is challenging [*Zhang et al.*, 1997; *Mao et al.*, 1999]. The uncertainty in tectonic inferences derived using different data depends on the uncertainties in both the data and model assumptions. In many cases [e.g., *Newman et al.*, 2001] it seems likely that estimates of the random (or formal) error expected from the data underestimate the systematic errors, so the overall uncertainty is dominated by the latter and thus larger than expected. This phenomenon, noted in other branches of science [*Henrion and Fischhoff*, 1986; *Ekeland*, 1993], can cause estimates of the same quantity derived from different techniques to differ by more than the assumed uncertainties for each technique. Hence it can be challenging to decide when results from different techniques are in fact sufficiently similar or different to draw tectonic conclusions.

Global plate motion models derived without space geodetic data offer a natural starting point for studying a boundary zone, in that they predict the net motion across it averaged over millions of years. The primary limitations of these models is that they are derived using data only at plate boundaries which are combined assuming that plates are rigid. Thus the predicted motion at a boundary can be biased by errors in data from both that and other boundaries, and by intraplate deformation either near that boundary or elsewhere in the global plate circuit [*Stein and Gordon*, 1984]. For example, because the convergence rate at subduction zones is not recorded directly by marine magnetic data, predictions of the convergence rate rely primarily on data from other boundary segments. Estimates of the direction of motion, which incorporate the slip vectors of trench earthquakes, can be biased if slivers of the overriding plate move independently from the rest of that plate, as discussed later.

Space geodetic data are the primary method of studying the kinematics of boundary zones on land, because they record both motions within boundary zones and global plate motions between plate interiors. They thus avoid the problems posed by global circuits and plate non-rigidity. The advances in space geodesy have been remarkable: only a few years ago, using GPS data to measure site motions to

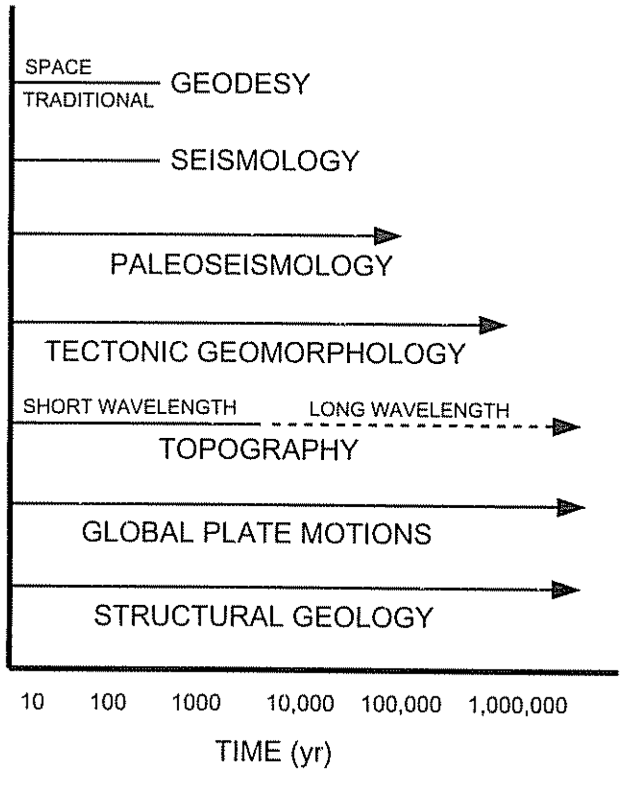

Figure 3. Schematic illustration of how plate boundary zone crustal deformation is observed by different techniques which sample various time scales.

sub-centimeter precision was a goal in itself. Now, these motions are combined with global GPS data to determine velocity fields over areas as large as a 1000-km wide plate boundary zone. These GPS-derived vectors are increasingly being made publicly available on the WWW, and results of various GPS studies are often combined with other space geodetic data to produce combined solutions for velocity fields. For example, a recent velocity field for the Western U.S. [*Bennett et al.*, this volume] combines data from eleven GPS networks with VLBI, SLR, and DORIS data. Space geodetic data from around the world are also being combined to develop global velocity models [*Argus and Heflin*, 1995; *Larson et al.*, 1997; *Kreemer et al.*, 2000; *Sella et al.*, 2002]. These velocities provide a powerful tool that can be used by various investigators with little consideration of the details of how they were acquired, as is done with earthquake focal mechanisms or plate motion rates from magnetic anomalies.

The precision of the inferred velocities is proportional to the uncertainty of successive position estimates divided by the time between them. Hence even if the uncertainty of individual measurements stays constant, the velocity uncertainty shrinks. Survey (campaign) mode GPS studies, in which antennas are set up over monuments for short periods and reoccupied later, give site positions which are repeatable to about a centimeter. Continuously recording GPS receivers permanently mounted at sites yield significantly more precise positions and thus velocities [*Heflin et al.*, 1992; *Mao et al.*, 1999; *Wernicke et al.*, 2000; *Hudnut et al.*, 2001; *Sella et al.*, 2002], albeit at higher cost. For either approach, the precision of the inferred velocity increases as data are acquired over longer time intervals [*Mao et al.*, 1999]. Thus a primary source of uncertainties in the velocity estimates are the short time series available at many sites. More accurate velocity estimates and denser spatial coverage make it possible to refine tectonic models and discriminate between alternatives.

Important issues arise in plate boundary zone studies when space geodetic data are compared to global plate motion models. Because space geodetic velocities for sites in plate interiors are generally similar to those predicted by plate motion models, it is interesting to explore cases where significant differences exist, to determine whether they reflect uncertainties and errors in one or both approaches, or instead reflect real differences in plate motions over different time scales. However, space geodetic velocities within boundary zones often differ from the predictions of plate motion models because geodetic velocities over a few years include the effects of transient elastic deformation associated with the cycle of strain accumulation and release in plate boundary earthquakes [*Savage*, 1983; *Scholz*, 1990], which is averaged out over the millions of years used in global plate motion models.

Seismology is another method of studying the directions and rates of the portion of the motion within plate boundary zones that occurs as slip in earthquakes. The rate of motion that occurs in earthquakes on discrete faults, typically the nominal plate boundary, is estimated from direct measurements of slip in earthquakes or the slip inferred from their seismic moments and the earthquake history [*e.g.*, *Brune*, 1968; *Kanamori*, 1977; *Engeln et al.*, 1986; *Kreemer et al.*, this volume]. Slip rates in broader regions containing a variety of faults, such as fold-and-thrust belts, are inferred by summation of seismic moment tensors [*Kostrov*, 1974; *Jackson and McKenzie*, 1988]. Such estimates require assumptions about the geometry of faulting, especially its depth, which is often poorly known. Another source of uncertainty in these estimates is that earthquakes on any given fault are infrequent, with recurrence times of hundreds or thousands of years. Given the short time period of instrumental seismology (since about 1900), only a small sample of the motion is well recorded [*McCaffrey*, 1997]. Historical records that give the dates of large past earthquakes span a somewhat longer period, but inferences about their locations and the motion in them have significant uncertainties, which in turn affect estimates of rates of motion. In many places it appears that only part of the total motion occurs in earthquakes, with the rest taking place aseismically either as aseismic motion between earthquakes or in aseismic afterslip [*Kanamori*, 1977; *Peterson and Seno*, 1984; *Jackson and McKenzie*, 1988; *Pacheco et al.*, 1993; *Bird et al.*, this volume; *Kreemer et al.*, this volume; *Langenhorst and Okal*, this volume]. Hence when seismological results appear discordant with those from global plate motion models and space geodesy, the discrepancy may reflect either the limitations of the earthquake record or the presence of aseismic deformation.

Paleoseismology, the analysis of geologic records of earthquakes, is used to extend the earthquake time history, and thus rates of motion, further into the past. Paleoseismic studies can identify prehistoric earthquakes, and give insight into their magnitudes. An important result of such studies is that earthquake recurrence intervals can be quite variable [e.g., *Sieh et al.*, 1989], so paleoseismic data improve estimates of rates of seismic slip from those using only the short time series of instrumental data [e.g., *Atwater*, 1992]. There may, however, be systematic discrepancies between paleoseismic results and those from instrumental seismology, with implications for combining paleoseismic data with geodetic, seismological, and plate motion data.

Paleoseismic studies often find that large ("characteristic") earthquakes occur more frequently than expected from the linear earthquake frequency - magnitude relation derived from instrumental data [*Youngs and Coppersmith*, 1985], although the reverse has also been noted [*Meghraoui et al.*, 2001]. However, seismological data for continental interiors [*Triep and Sykes*, 1997] show that large earthquakes are less frequent than expected from the linear relation, presumably because of the finite depth available for faulting. Thus differences between seismological and paleoseismic results may reflect the effects of small sampling in which the frequency-magnitude relation is better estimated from the more common small earthquakes than the infrequent larger ones [*Howell*, 1985]. Alternatively, the discrepancy may be due to difficulties in assessing magnitudes using paleoseismology, or may be real.

Structural geology and tectonic geomorphology also constrain rates and directions of motion. We can view these techniques as a continuum with paleoseismology, which focus on the net geologic effects over longer periods for which the effects of individual earthquakes and aseismic deformation are summed. For example, geologic estimates of displacements along faults span many earthquake cycles and so give an average rate of motion [e.g., *Sieh and Jahns*, 1984; *Holt et al.*, 2000] that can be compared to data from plate motion, seismological, and geodetic studies. Structural data can also be used to derive two-dimensional long-term velocity fields in thrust belts for comparison with the short-term geodetic data [*Hindle et al.*, 2002; *Hindle and Kley*, this volume]. Although the uncertainties due to field measurements, age assignments, and different techniques are difficult to characterize, structural and geomorphological approaches provide valuable comparisons with the other methods. Additional data can be obtained from paleomagnetic studies, which show block rotations in plate boundary zones [e.g., *Courtillot et al.*, 1984; *Beck et al.*, 1994; *Lamb*, 2001].

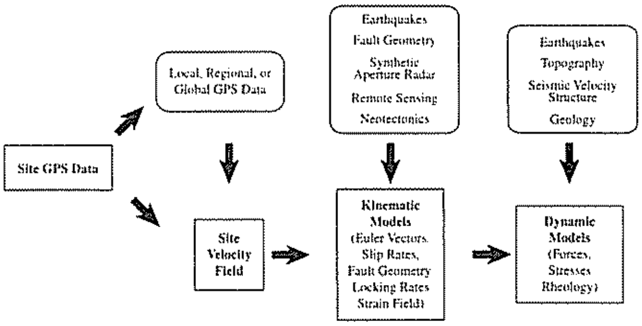

Figure 4. Schematic illustration of some of the ways various types of data can be integrated in plate boundary zone studies.

Topography can yield important information for boundary zone studies. Even though it does not directly give rates of motion, it is created in part by the horizontal crustal motions in boundary zones and thus reflects them. For example, the width of basins can constrain the extension history [*Acton et al.*, 1991; *Dokka and Macaluso*, 2001] and, as discussed later, the height and form of mountain ranges can be used to as a input to or constraint on models of plate boundary zone dynamics [e.g., *Jones et al.*, 1996; *Flesch et al.*, 2000; *Liu et al.*, 2000, this volume; *Coblentz and Stüwe*, this volume].

Plate boundary zones are being studied using various combinations of these data. Figure 4 illustrates one such approach, beginning with GPS data. First, kinematic models describing plate motions, fault geometry, or deformation fields can be developed by combining GPS results with data including earthquake locations and mechanisms, geological fault maps, or remote sensing data. Next, dynamic models of the forces, stresses, and rheology can be developed, that can incorporate data such as topography and seismic velocity structure. Finally, models can be extended backward in time using geologic information including plate motion models based on magnetic anomalies, faulting and uplift histories. Hence the kinematic data provide valuable constraints on models of plate driving forces, as illustrated by *Hamilton* [this volume], *Anderson* [this volume], and *Zatman and Richards* [this volume].

KINEMATICS AND DIMENSIONS OF PLATE BOUNDARY ZONES

Whether to regard a region as "interplate" or "intraplate" often depends on the definitions we choose for "plates" and "plate boundaries", as discussed by *Stein et al.* [this volume] and *Li et al.* [this volume]. Increasingly, this choice is made by using the techniques just reviewed to study crustal motions along plate boundaries and define the extent of the zones over which the relative plate motion is distributed. These zones are then regarded as the boundary zone, and distinguished from the stable plate interior. Within these boundary zones, studies are investigating whether the motions are better described as varying smoothly or as taken up by a few microplates or blocks, and how the large-scale motions of the major bounding plates relate to the boundary zone deformation.

Figure 5 illustrates the type of data becoming available, in this case for motion across the North America-Pacific boundary zone in western North America. The idea of this region as a wide boundary between rigid Pacific and North American plates was advocated by *Atwater* [1970] following pre-plate tectonic concepts of *Carey* [1958], *Wise*

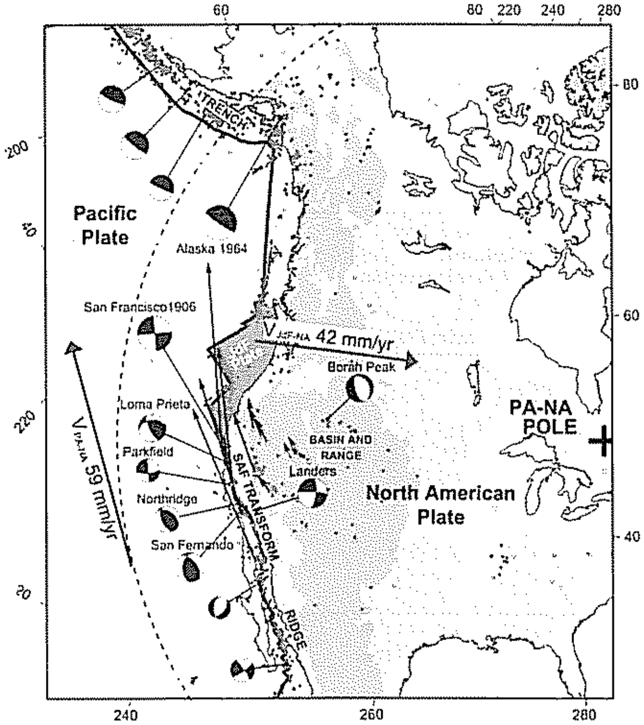

Figure 5. Geometry and focal mechanisms for a portion of the North America-Pacific boundary zone that also includes the small Juan de Fuca (JdF) plate. Dashed line shows a small circle, the direction of plate motion, about the Pacific-North America Euler pole. This small circle is further from the pole than the San Andreas fault, so motion along it is faster. The variation in the boundary type along its length from extension, to transform, to convergence, is shown by the focal mechanisms. The diffuse nature of the boundary zone is shown by seismicity (dots), focal mechanisms, topography (1000 m contour shown), and vectors showing the motion of GPS and VLBI sites (squares) with respect to stable North America. The velocity scale is shown by the plate motion arrows; some site motion vectors are too small to be seen. Modified from *Stein and Klosko* [2002]

[1963], and *Hamilton and Myers* [1966], and was generally accepted based on the broad distribution of seismicity, elevated topography, volcanism, and active faulting. However, it was unclear whether the difference between the ~36 mm/yr motion on the San Andreas fault inferred from geologic data over the past several thousands of years [*Prescott et al.*, 1981; *Sieh and Jahns*, 1984] and the ~48 mm/yr of Pacific-North America plate motion predicted by models averaging data over the past 3 Ma [*DeMets et al*, *1987*] reflected a difference between short and long term rates, or is accommodated elsewhere in the boundary zone. Space geodetic data [*Clark et al.*, 1987; *Dixon et al*, 1991; *Shen et al.*, 1997; *DeMets and Dixon*, 1999; *Bennett et al.*, 1999, this volume] have resolved this issue. The net motion across the zone is essentially that predicted by global plate motion model NUVEL-1A, and the motion that does not occur along the San Andreas fault is distributed across the boundary zone. The pattern of motion is complicated [*Dokka and Travis*, 1990; *Hearn and Humphreys*, 1998]. Geodetic data show that the boundary zone includes the Sierra Nevada microplate, a relatively rigid block east of the San Andreas fault system [*Wright*, 1976; *Argus and Gordon*, 1991; *Dixon et al.*, 2000; *Miller et al.*, 2001a] and the zone of Basin and Range extension [*Thatcher et al.*, 1999; *Wernicke et al.*, 2000].

The seismicity and earthquake mechanisms also show the extent and complexity of the boundary zone. In addition to primarily strike-slip motion on the San Andreas itself, thrust faulting occurs in events like the 1971 San Fernando and 1994 Northridge earthquakes and normal faulting is seen due to the regional extension in the Basin and Range province. The boundary zone is further complicated by volcanism including Long Valley caldera in eastern California and Yellowstone hot spot, which also have associated seismicity. Thus large (magnitude 7) earthquakes as far east as Nevada and Utah result from Pacific – North American plate motion. Profiles of site velocity in the plate motion direction (Figure 6) thus vary across the boundary zone from essentially zero within one plate's interior to the full plate motion rate within the other's interior [e.g., *Bennett et al.*, 1999].

Space geodesy is being combined with earthquake studies to study subduction and associated faulting in Alaska, the longest and most seismically active portion of the plate boundary [*Larson and Lisowski*, 1994; *Cohen*, 1996; *Sauber et al*, 1997; *Freymueller et al.*, 2000]. Similarly, space geodesy is being used to study how the subducting Juan de Fuca plate deforms the overriding North American plate [*Dragert et al.*, 2001; *McCaffrey et al.*, 2000; *Miller et al.*, 2001b; *McCaffrey*, this volume]. The GPS data are particularly valuable here because the last large interplate thrust earthquake occurred in 1700 AD, prior to the invention of the seismometer, so the subduction geometry is poorly known.

Western North America illustrates that plate boundary zones are significantly broader than the region over which transient deformation occurs due to the elastic strain accumulation between major earthquakes on an individual fault [*Mavko*, 1981] that can be treated as the nominal plate boundary. Figure 6c shows a GPS velocity profile across the much-photographed Carrizo Plain segment of the San Andreas fault, which is consistent with the arctangent curve expected for interseismic displacement in an elastic rebound model of the seismic cycle on an infinite strike-slip fault at a plate boundary [*Savage and Burford*, 1973]. Between

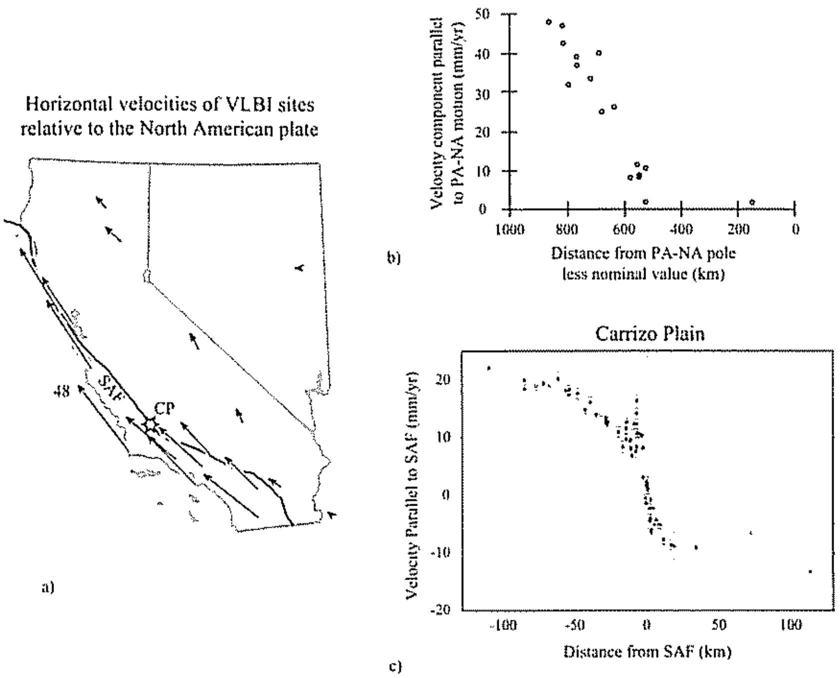

Figure 6. Variation in motion of space geodetic sites across parts of the Pacific-North America boundary zone. a) Horizontal velocities of sites in California, Nevada, and Arizona relative to stable North America. The velocity of the southwesternmost site nearly equals the predicted 48 mm/yr velocity of the Pacific plate relative to the North American plate b) Data from (a) showing component of motion tangent to small circles centered on the Pacific-North America Euler pole versus angular distance from that pole. Velocities increase with distance from the Euler pole, and most of the variation occurs across the San Andreas fault [*Gordon and Stein*, 1992] c) GPS data showing the rate of fault-parallel motion, about 35 mm/yr, across the Carrizo Plain segment of the San Andreas fault whose location is shown by the star [*Z.-K. Shen, personal communication*, 2000].

earthquakes motion occurs away from the locked fault, but near it interseismic strain and displacement build up until they are released by the coseismic motion, causing motion at the fault to "catch up" with the plate motion. Most of the strain accumulation occurs within about 50 km of the fault, consistent with a fault locked to about 20 km depth. The total motion across this fault segment is about 35 mm/yr, with the remainder of the ~50 mm/yr of plate motion spread over the much broader boundary zone (Figure 6a,b). For a broad boundary zone containing a number of parallel strike-slip faults, the net effect of the faults gives a similar-appearing broad velocity profile, as shown for Southern California by *Bourne et al.* [1998].

Detailed views of the velocity field are also being acquired for many other plate boundary zones. For example, space geodetic data (Figure 7) show the kinematics of the broad collision between India and Eurasia. This area is the present type example of mountain building by continental collision, which has produced a boundary zone extending thousands of kilometers from the nominal plate boundary at the Himalayan front. One primary issue is the relative importance of two mechanisms, crustal thickening and lateral extrusion of crust, in taking up the convergence. Estimates of the extrusion fraction vary from 10-50% [*Molnar et al.*, 1987; *Avouac and Tapponnier*, 1993], and determination of this quantity has important consequences for models of the mechanics of convergence [*Houseman and England*, 1986; *England and Houseman*, 1986; *England and Molnar*, 1997]. GPS data directly measure rates and directions of motion in different portions of the boundary zone, which could previously only be estimated indirectly. They illustrate overall plate convergence at a rate slower than NUVEL-1A's prediction [*Chen et al.*, 2000; *Shen et al.*, 2000], shortening across the locked Himalayan front which gives rise to large destructive earthquakes [*Bilham et al.*, 1997; *Larson et al.*, 1999], E-W extension in Tibet, and eastward escape of continental fragments [*King et al.*, 1997]. The extent of the collision is illustrated by the data showing that the Tien Shan intracontinental mountain belt, 1000-2000 km north of the Himalaya, accommodates almost half the net plate convergence in the western part of the zone [*Abdrakhmatov et al*, 1996]. Thus efforts to understand the collision trend naturally into studies of kinematics of southeast Asia [*Wolpersdorf et al.*, 1998; *Rangin et al.*, 1999; *Chamot-Rooke and Le Pichon*, 1999] and far eastern Asia [*Kato et al.*, 1998a,b; *Calais et al*, 1998; *Heki et al.*, 1999; *Park et al.*, this volume].

A related and long-standing class of problems relate to the plate geometry and motions in northeastern Asia, one of the few remaining places where even the approximate boundaries of the major plates remain in question. Although the North America-Eurasia boundary can be traced along the Mid-Atlantic Ridge to the Arctic (Nansen) Ridge from seismicity, the seismicity decreases and becomes more diffuse once it reaches the Eurasian continent. *Morgan* [1968] hence noted that "the boundaries in Siberia and central Asia are very uncertain" and that "additional subblocks may be required." Despite subsequent studies, the plate geometry has not been conclusively resolved, and it is unclear where the Eurasia-North America boundary is, whether the East Siberia-Sea of Okhotsk-northern Japan region should be treated as part of the North American plate or as a separate Okhotsk plate, and whether the region west of Japan and Sakhalin in Siberia and north China should be treated as part of Eurasia or as a distinct Amurian plate [*Chapman and Solomon*, 1976; *Zonenshain and Savostin*, 1981; *Cook et al.*, 1986; *Wei and Seno*, 1998]. *Seno et al.* [1996] analyzed earthquake and plate motion data from the area, and concluded that "most future insight into the issue of the relative motions will come from the increasing availability of relevant space geodetic data." Initial results from GPS sites in Siberia [*Kogan et al.*, 2000], Japan [*Takahashi et al.*, 1999], and east Asia [e.g. *Calais et al.*, 1998; *Heki et al.*, 1999], show significant progress on these issues.

PLATE INTERIORS

The converse of the process of using geodetic site velocities to define plate boundary zones is using them to define stable plate interiors. Although crustal motions, earthquakes, faults, volcanoes, and topography are physical entities, "plates" are human constructs. Initially, plates were defined as regions within which deformation, as measured by active faulting, earthquakes, or other criteria, seemed slower than in regions regarded as plate boundaries. This approach allowed the major plates to be identified and their relative motions calculated using methods that assumed plate rigidity. However, the assumption of rigidity was difficult to quantify except via the nonclosure of plate motion circuits. Thus whether to regard deformation such as earthquakes as "intraplate", as opposed to occurring on a slowly-moving plate boundary, was in some cases unclear.

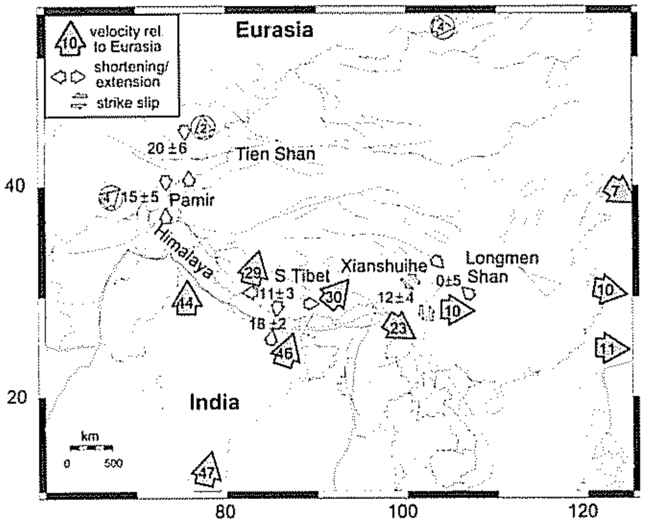

Figure 7. Summary of crustal motions determined using space geodesy in the India-Eurasia plate collision zone. Large arrows indicate velocities in mm/yr relative to Eurasia. Arrows in circles show velocities with no significant motion with respect to Eurasia. Small arrows show local relative deformation. Thick grey lines are nominal plate boundaries. Modfied from *Larson et al.* [1999].

As global plate motion data improved, it has become increasingly feasible to test whether data are statistically better fit by assuming the existence of two distinct plates [*Stein and Gordon*, 1984; *Gordon et al.*, 1987]. Such analyses showed that North and South America [*Stein and Gordon*, 1984], India and Australia [*Wiens et al.*, 1985], Rivera and Cocos [*DeMets and Stein*, 1990], and Nubia and Somalia [*Chu and Gordon*, 1999] should be regarded as distinct plates, often with seismicity along their boundaries, rather than single plates with distinct zones of intraplate seismicity. In such cases, a two-plate description is useful in that an Euler vector can be derived and used to describe the motion of the two plates.

Eventually, however, deformation becomes slow and diffuse enough that it is more usefully regarded as intraplate. Space geodesy is ideal for addressing this issue because it can measure motions of a few mm/yr. As a result, there has been considerable interest in using space geodesy to quantify the rigidity of major plates and investigate how deviations from plate rigidity give rise to intraplate deformation and earthquakes. This is an interesting scientific issue with societal implications because intraplate earthquakes can be quite destructive, but are hard to study because they are much rarer than those on plate boundaries.

The first step to estimate the rigidity of a plate using space-based geodesy is to find the motions of sites within it, and compare these motions to those predicted assuming that the plate is rigid and so can be described by a single Euler vector [*Dixon et al.*, 1996; *Newman et al.*, 1999; *Kogan et al.*, 2000; *Noquet et al.*, 2000; *Sella et al.*, 2002; *Beavan et al.*, 2002]. For example, application of this approach to eastern North America shows that the plate is quite rigid. Successive studies (Table 1) using the steadily increasing lengths of data from the growing number of continuous GPS sites, which provide daily positions and hence very precise velocities, show that the root-mean-square misfit of site velocities to those expected from a single Euler vector if the plate were perfectly rigid is now less than 1 mm/yr.

The misfit is strikingly small, given that it is a measure of the combined effects of intraplate deformation due to tectonics and glacial isostatic adjustment, uncertainties in the positions of geodetic monuments due to the GPS techniques, and local motion of the geodetic monuments. The result seems plausible because similar values emerge from VLBI studies [*Argus and Gordon*, 1996]. Hence sites moving faster with respect to the stable interior of the plate than a specified rate, perhaps 2-3 mm/yr, can be viewed as within the boundary zone. Figure 8 illustrates this concept using velocities of GPS sites with respect to the Pacific. Velocities in the essentially rigid plate interior move coherently about the Euler pole, whereas those in the plate boundary zones deviate from this motion.

The next step is to use GPS velocities to estimate the intraplate strain rate field in various areas and compare it to present-day seismicity, paleoseismicity, and geologic data. This is in principle straightforward; the strain field can be derived by forming least-squares estimates of the velocity gradient in various regions. Application of this technique in the Andes [*Klosko et al.*, this volume] yields strain rate tensors of magnitudes about 5 x 10^{-8} 1/yr, showing the shortening within the South American plate that is building the Andes.

However, when we applied this technique to eastern North America two years ago [*Klosko et al.*, 2000], we found no statistically significant deformation at a sensitivity at about 10^{-9} 1/yr. Similarly, *Gan and Prescott* [2001] found no regions with statistically significant strain rates. Hence, as we might expect, the intraplate strain rate field is low in magnitude. It is worth noting that 1 mm/yr of motion across 100 km would yield 10^{-8} 1/yr, about 1/5 of that in the Andes, and 1 mm/yr across 1000 km would yield 10^{-9} 1/yr. As a result, GPS studies have not yet detected intraplate strain accumulation, but show that intraplate strain accumulation rates are slow, which yields useful insight into seismic hazards [*Newman et al.*, 1999, 2001; *Gan and Prescott*, 2001; *Stuart*, 2001]. Nonetheless, given the rapidly growing number of continuous GPS sites and the longer spans of GPS data, it seems likely that the strain accumulation signal will soon "climb" above the noise and provide a valuable datum for investigation of intraplate tectonics.

Table 1. Estimates of the rigidity of eastern North America

Study	Number of GPS Sites	rms misfit (mm/yr)
Dixon et al. [1996]	8	1.3
Newman et al. [1999]	16	1.0
Klosko et al. [2000]	54	0.93
Sella et al. [2002]	64	0.86

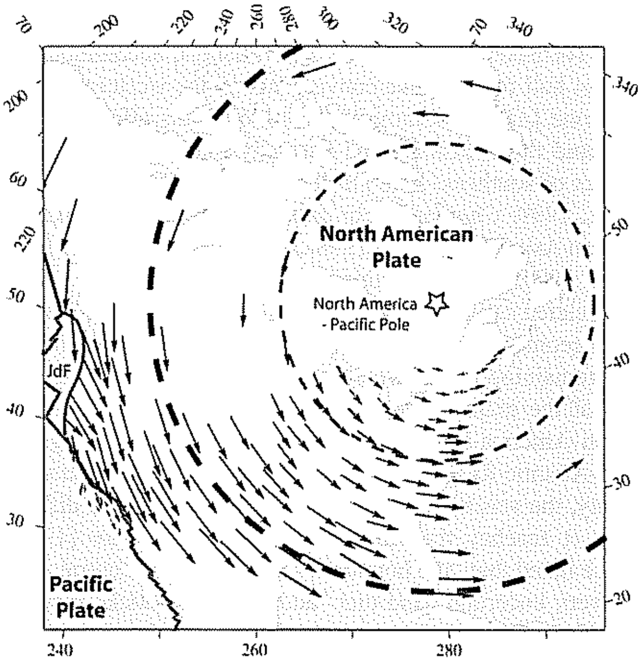

Figure 8. GPS site velocities with respect to the Pacific plate. Sites within the essentially rigid North American plate interior rotate coherently about the North American-Pacific pole [*Sella et al.*, 2002]. Linear velocities vary as the sine of the angular distance from the pole, and directions follow small circles, like those shown, about it. Site motions within the plate boundary zone deviate significantly from the rigid plate motion.

MICROPLATES

In many cases, a region within a boundary zone between two major plates moves as a coherent rigid block or microplate (the terms are equivalent) with little internal deformation Microplates have long been recognized in boundary zones between oceanic plates, because their relative motions are recorded by marine magnetic anomalies and so can be seen along present plate boundaries such as the East Pacific Rise near Easter Island [*Engeln and Stein*, 1984; *Hey et al.*, 1985; *Engeln et al.*, 1988] and the Pacific - Nazca - Antarctic triple junction [*Anderson-Fontana et al.*, 1986; *Bird et al.*, 1998]. Seismicity can show that deformation is concentrated at their boundaries, and analysis of the microplate's motion can show that it obeys rigid plate kinematics described by rotations about Euler poles Characteristic patterns of magnetic anomalies similar to those observed for present-day microplates are preserved in the sea floor, showing that microplates were common in the past [*Cande et al.*, 1982; *Mammerickx et al.*, 1988; *Tamaki and Larson*, 1988; *Fullerton et al.*, 1989]. Without the benefit of marine magnetic anomalies, the boundaries between continental blocks are more difficult to identify, but blocks have been identified using paleomagnetic and structural data [*Achache et al.*, 1984; *Brown and Golombek*, 1985; *Garfunkel and Ron*, 1985; *Acton et al.*, 1991].

Space geodetic data greatly simplify identification and study of continental microplates, as illustrated by the Sierra Nevada block within the Pacific - North America boundary zone, as discussed earlier. Often the data are sufficient to estimate an Euler vector describing the microplate's motion relative to the major plates. Figure 9 illustrates a model derived from GPS site velocities in which an Oregon block moves relative to both North America and the Juan de Fuca plate.

This example illustrates a common feature of microplate systems, that the Euler pole for rotation of the microplate with respect to an adjacent major plate is close to the microplate [*Engeln et al.*, 1988; *Schouten et al.*, 1993]. Figure 10 shows how this proximity occurs for a microplate bounded on the east and west by spreading centers, such that the east ridge propagates northwards. In this example the pole for the relative motion between the major plates is far away, so the rate of relative motion between them (which varies as the sine of the angle between a site and the pole) varies slowly along this part of the boundary. However because motion varies rapidly along the microplate's boundary with the eastern major plate, the pole for the microplate's motion with respect to this plate must be nearby. For the rate of this motion to be comparable in places to that between the major plates the rotation rate, which is the magnitude of the Euler vector, must be larger than that for motion between the major plates. By vector addition, the Euler vector for the microplate's motion with respect to the western major plate must also be large and have a nearby pole. Similar geometries can occur in other situations where plate motions vary rapidly over short distances.

Identification or confirmation of microplates within plate boundary zones is helping resolve long-standing issues of plate boundary kinematics. A particularly striking example is the set of complex boundary zones bordering the Caribbean plate. In global models like NUVEL-1, the motion of the Caribbean plate was expected to be the worst known of any major plate, because data were sparse. A magnetic anomaly rate can be measured at only one place, the Cayman spreading center, and the Caribbean's boundaries with the North and South American plates are so complicated that it is unclear whether earthquake slip vectors

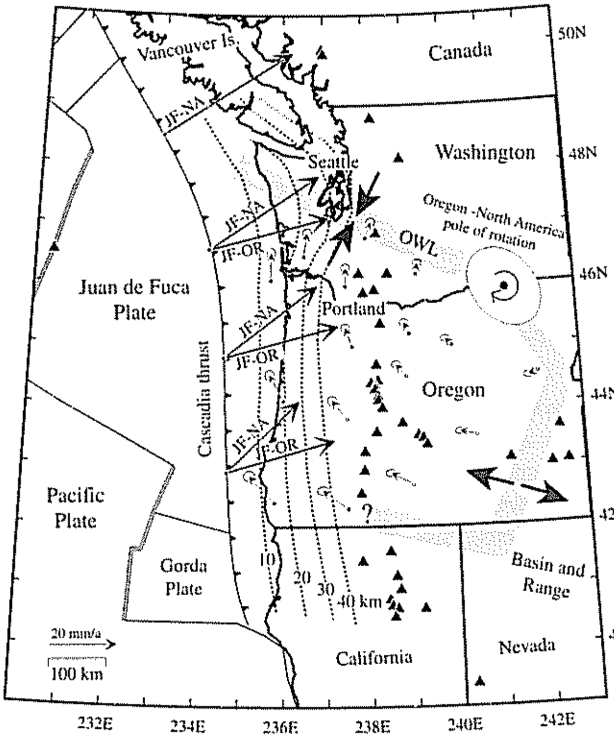

Figure 9. A model in which a microplate containing most of Oregon and SW Washington (OR) rotates about a nearby pole in response to Basin and Range extension. The boundary geometry predicts contraction along the Olympic-Wallowa lineament (OWL), extension in the Basin and Range, convergence at the Cascadia subduction thrust, whereas the location of the southern boundary is not clear. The rotation makes the subduction direction less oblique to the trench than Juan de Fuca - North America (JF-NA) motion. Shown are active volcanoes (triangles) and depth contours of the top of the subducting Juan de Fuca plate (dashed lines). Large arrows show convergence of the Juan de Fuca plate with both North America (JF-NA) and the rotating Oregon block (JF-OR) at the Cascadia deformation front. Ellipse in NE Oregon shows pole and 1-s uncertainty for rotation of OR relative to NA. Small arrows show velocities and 1-s uncertainties relative to NA predicted by this pole [Modified from *McCaffrey et al.*, 2000].

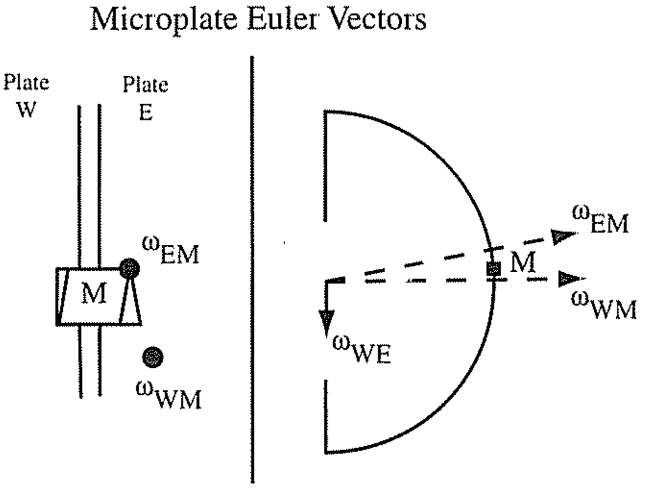

Figure 10. Geometric constraints on the Euler vectors for a three-plate system with a microplate ("M") between western ("W") and eastern ("E") major plates. The Euler vector for the major plates ω_{WE} is far away, so their relative motion varies slowly along this part of the boundary, whereas those for the microplate are nearby and larger in magnitude, yielding relative motions which vary rapidly and are comparable in places to the relative motion between the major plates [*Engeln et al*, 1988].

reflect overall plate motion or block motions in a complex boundary zone. As a result, plate motion models incorporating magnetic anomaly data from the Cayman spreading center [*Jordan*, 1975; *Stein et al.*, 1988; *DeMets et al.*, 1990, 1994] predicted less than 20 mm/yr of North America-Caribbean motion, whereas *Sykes et al.* [1982] and *Deng and Sykes* [1995] inferred a higher convergence rate from the length of the Lesser Antilles Benioff zone. GPS data find a velocity intermediate between the models [*Dixon et al.*, 1998; *DeMets et al.*, 2000; *Jansma et al.*, 2000; *Sella et al.*, 2002], consistent with the proposed Gonave microplate within the North America - Caribbean plate boundary zone, east of the Cayman spreading center [*Rosencrantz and Mann*, 1991; *Mann et al.* 1995; *Leroy*, 2000]. Hence spreading at the Cayman spreading center is only part of the relative plate motion, with another portion taken up south of the nominal plate boundary on faults including the Enriquillo fault zone in the Dominican Republic. This geometry has implications for earthquake hazards.

Microplate motion is also observed in central and northern South America where the Nazca, Cocos, Caribbean and South American plates interact. Although there is general agreement that plate convergence occurs, the locations of the boundaries have long been a subject of controversy [e.g. *Dewey*, 1972] and contributed to the difficulty in defining Caribbean plate motions. GPS measurements are improving our understanding of the complex tectonics [*Freymueller et al.*, 1993; *Kellogg et al.*, 1995, 2001; *Weber et al.*, 2001; *Perez et al.*, 2001]. It appears that some of the complexity reflects the presence of a distinct North Andes block, between the Caribbean and South American plates, that moves northward relative to stable South America.

GPS data show microplate motion within the Mediterranean collision zone involving the Nubian (West African), Arabian, and Eurasian plates, which has long been regarded as a complex set of blocks [*McKenzie*, 1972]. Although efforts to unscramble the motion using earthquake focal mechanisms and other data have proved challenging, GPS data provide a compelling picture (Figure 11). Studies in the western Mediterranean, [*Reilinger et al.*, 1997; *McClusky et al.*, 2000] show the northward motion of Arabia and its effects. Anatolia rotates as a microplate about a pole near the Sinai peninsula, as it is "squeezed" westward between Eurasia and northward-moving Arabia, a situation that has been likened to a melon seed squeezed between a thumb and forefinger. Motion across the North Anatolian fault, about 25 mm/yr, gives rise to large strike-slip earthquakes like the 1999 $M_w = 7.4$ Izmit earthquake. Increasing velocities toward the Hellenic arc, where Nubia subducts below Crete and Greece, show that western Anatolia and the Aegean region are under extension, consistent with the normal fault mechanisms observed. Anatolia may be deforming and being "pulled" toward the arc by the foundering Nubian plate or mantle convection in the Aegean region [*Sonder and England*, 1989; *Royden*, 1993]. In contrast, eastern Turkey is being driven against Eurasia, causing compressive strain that is accommodated by strike-slip faulting in eastern Turkey and compression across the Caucasus. To the south, motion along the Dead Sea transform reflects the relative motion of the Sinai microplate with respect to Arabia [*Joffe and Garfunkel*, 1987; *Le Pichon and Gaulier*, 1988; Badawy and Horváth, 1999; *Mascle et al.*, 2000; *Pe'eri et al.*, 2002; *Al-Zoubi*, this volume]. In addition *Devoti et al.* [this volume] combine GPS, SLR, and VLBI to show how the Adriatic region collides with Eurasia, and *Grenerczy* [this volume] uses GPS and focal mechanism data to characterize the effects of this collision on central Europe.

Microplate models are under discussion for many other areas. As discussed shortly, portions of the overriding plate at subduction zones often appear to move as a coherent sliver block between two major plates. GPS data have been interpreted as showing either a smooth distribution of extension across the Basin and Range [*Bennett et al.*, 1999] or extension concentrated adjacent to the Sierra Nevada block, such that western Utah and eastern Nevada act rigidly [*Thatcher et al.*, 1999]. *Stein et al.* [this volume] suggest that

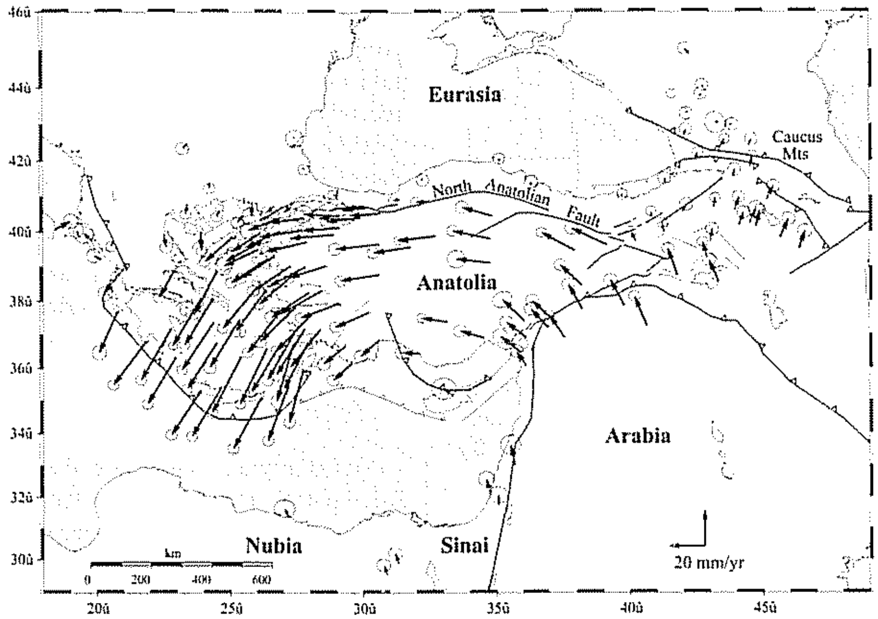

Figure 11. GPS observations of motion relative to Eurasia for a portion of the Africa-Arabia-Eurasia plate collision zone. Note the motion of the Anatolian microplate, strike-slip along the North Anatolian fault, extension in western Anatolia and the Aegean region, and compression in the Caucasus mountains. Modified from *McClusky et al.* [2000].

a microplate within the broad India - Arabia - Eurasia triple junction region is breaking off from the Indian plate due to stresses associated with the Himalayan collision [*Li et al.*, this volume]. In this model, the microplate reflects the evolution of a complex boundary zone, as observed elsewhere, especially near triple junctions.

CONVERGENT BOUNDARIES

Interesting complexities are associated with convergent plate boundary zones, where the total convergence can be divided in various ways between motion at the trench, the primary plate interface, and deformation of the overriding plate. These complexities are illustrated in Figure 12 for the boundary zone where the oceanic Nazca plate subducts beneath continental South America. GPS site velocities, focal mechanisms, and elevated topography vary from the stable interior of the Nazca plate, across the Peru-Chile trench to the coastal forearc, across the high Altiplano and foreland thrust belt, and into the stable interior of the South American continent. The GPS velocities, which are shown relative to stable South America, would be zero if the South American plate were rigid and all motion occurred at the trench plate boundary. Instead, the site velocities are highest near the coast and decrease toward the interior of South America. The large motions near the coast are due to locking at the subduction interface, which causes the cycle of elastic strain accumulation and release in large interplate thrust earthquakes (Figure 13) like those whose focal mechanisms are shown. The GPS data show that motion extends much further inland than predicted by models of the seismic cycle, in which motion has largely decayed by a distance equal to twice that between the trench and locked fault end. The motions further inland across the subAndean foreland fold and thrust belt appear to reflect crustal shortening, also shown by thrust fault mechanisms, which is thought to have primarily responsible for building the Andes over time [*Dewey and Lamb*, 1992; *Allmendinger et al.*, 1997].

Such data are being used to model the kinematics of the present convergence and shortening [*Leffler et al.*, 1997; *Norabuena et al.*, 1998; *Angermann et al.*, 1999; *Bevis et al.*, 2001], combined with plate motion data to explore the time-variation in the subduction history [*Norabuena et al.*, 1999], and compared to earthquake data to examine the roles of seismic and aseismic deformation [*Klosko et al.*, this volume]. The GPS data are also being combined with geological data to study the history of crustal shortening [*Lamb*, 2000; *Hindle et al.*, 2002; *Hindle and Kley*, this volume], model the joint roles of tectonics and climate on the evolution of the Andean foreland thrust belt [*Horton*, 1999],

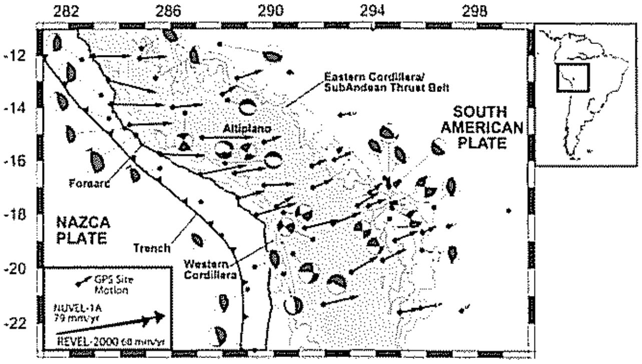

Figure 12. GPS site velocities relative to stable South America and selected earthquake mechanisms in the boundary zone. Rate scale is given by the NUVEL-1A and REVEL-2000 [*Sella et al.*, 2002] vectors. Modified from *Klosko et al.* [this volume].

and constrain models of the dynamics of convergence [*Liu et al.*, 2000; this volume]. Similar studies are being conducted at other convergent boundaries. We next discuss three aspects of such studies.

Slip Partitioning

Major thrust earthquakes at the interface between subducting and overriding plates characterize the subduction. In most cases, their focal mechanisms show slip toward the trench, approximately in the convergence direction predicted by global plate motion models or space-based geodesy. However, in some cases when the plate motion is oblique to the trench, a forearc sliver moves separately from the overriding plate (Figure 14). This effect, called slip partitioning, makes earthquake slip vectors at the trench trend between the trench-normal direction and the predicted convergence direction [*Jarrard*, 1986; *Ekstrom and Engdahl*, 1989; *DeMets et al.*, 1990; *DeMets and Stein*, 1990; *McCaffrey*, 1991, 1992, this volume] and causes strike-slip motion between the forearc and the stable interior of the overriding plate. In the limiting case of pure slip partitioning, pure thrust faulting would occur at the trench, and all the oblique motion would be accommodated by trench-parallel strike-slip.

This process is of interest for several reasons. The extent of slip partitioning may reflect the force balance between the thrust and strike-slip faults, and hence provide information on the physics of both interfaces [*McCaffrey*, 1992]. Moreover, slip partitioning may bias estimates of plate motions, because slip vectors may not represent motion between the major plates. Hence discrepancies between space geodetic data and a plate motion model may in part result from slip partitioning elsewhere in the global plate circuit [*DeMets et al.*, 1990].

Because slip partitioning was previously inferred from comparisons of slip vectors to the predictions of plate motion models its significance was difficult to assess until recently. However, GPS data can directly detect such sliver motions, and identify deformation within forearcs. Such data is becoming available from regions including Indonesia [*Prawirodirdjo et al.*, 1997; *Stevens et al.*, this volume], Costa Rica [*Lundgren et al.*, 1999] and Cascadia (Figure 9). Such observations are being used to model the regional kinematics and constrain numerical models of slip partitioning [*McCaffrey*, this volume].

Interplate Coupling

An underlying issue is how the physics of the plate interface controls the fractions of the plate motion taken up by various processes at the interface and within the overriding plate. These ideas are discussed in terms of seismic coupling, the fraction of plate motion that occurs by trench earthquakes, via two end members: coupled Chilean-type

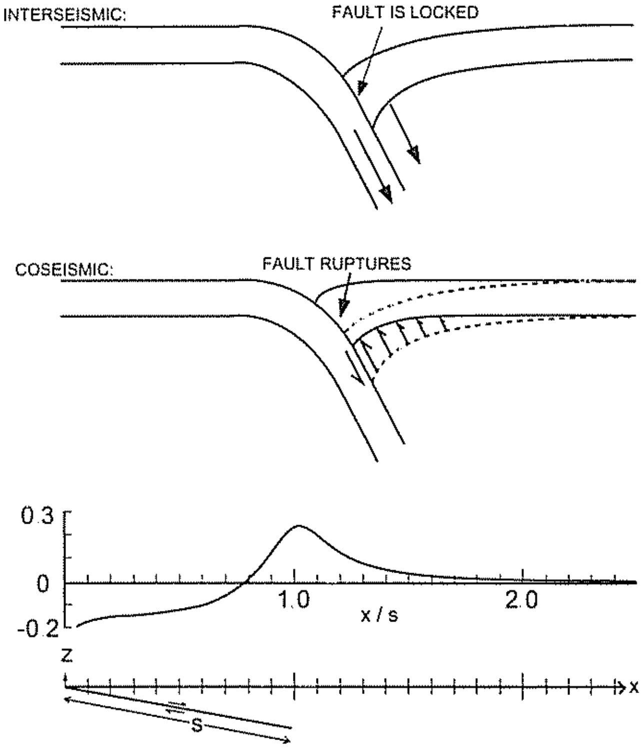

Figure 13. Top: Two stages in the earthquake cycle at a subduction zone [*Stein and Wysession*, 2002]. Bottom: Predicted interseismic vertical motion due to a locked fault at a subduction zone. The vertical motion is normalized by the locked plate convergence rate, and the horizontal distance is normalized by the distance between the trench and the end of the locked fault [*Savage*, 1983].

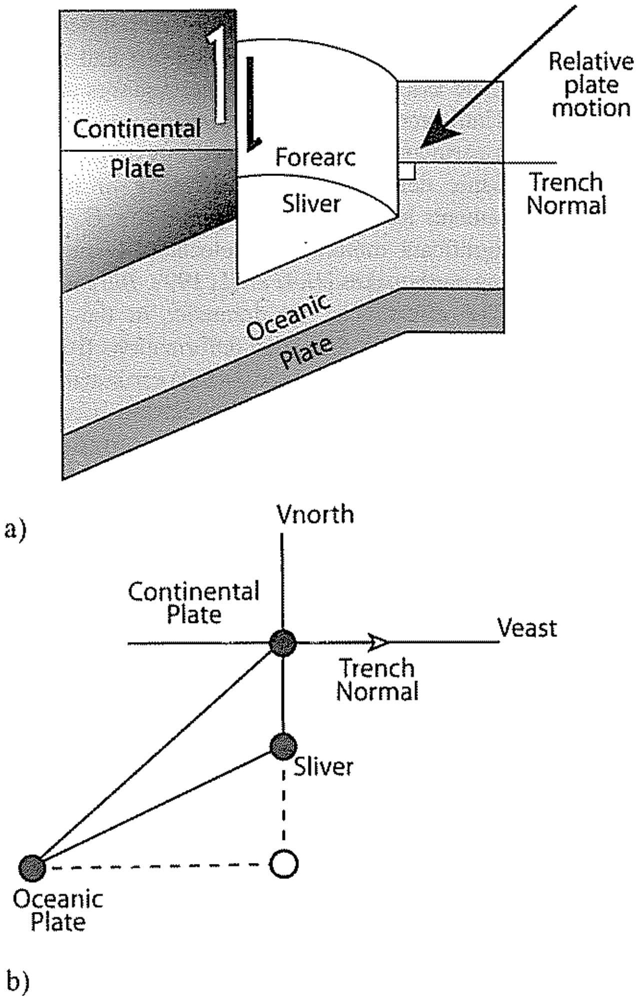

Figure 14. a) Schematic illustration of forearc sliver motion when convergence is oblique. b) Velocity space representation of slip partitioning, as shown in (a). Sliver motion is shown by solid circle and lines for partial partitioning, open circle and dashed lines for total partitioning.

zones with large earthquakes and uncoupled Mariana-style zones with largely aseismic subduction [*Kanamori*, 1977]. However, seismic coupling has proved difficult to measure and interpret

Prior to the advent of GPS data, seismic coupling was estimated only from earthquake histories. Slip in past earthquakes is estimated from their seismic moment and fault area, which are measured for instrumentally recorded (approximately post-1900) earthquakes and inferred for earlier earthquakes from historical records. The net slip over a certain time interval is then inferred from the recurrence history of past earthquakes, and a global plate motion model is used to assess the seismic coupling. This process has several limitations. First, the uncertainties in estimating source parameters of earthquakes from historical data are considerable. Second, the seismic cycle is typically longer than the instrumental record [*McCaffrey*, 1997], and the size and recurrence interval of earthquakes on a given trench segment can be quite variable [*Stein et al.*, 1986; *Thatcher*, 1990]. Third, some large earthquakes have significant slow slip associated with the main event, on time scales of days to years [*Kanamori and Cipar*, 1974; *Cifuentes and Silver*, 1989; *Barrientos*, 1995; *Heki et al.*, 1997]. The seismic moment released by these processes, termed slow or silent earthquakes or afterslip, is associated with earthquakes but is not included in conventional seismic moment calculations, and hence may produce a spurious seismic slip deficit. Hence it is difficult to assess whether an apparent seismic slip deficit indicates a seismic gap where large earthquakes are overdue, or that much of the interplate motion occurs aseismically [*Stein*, 1992].

Estimation of aseismic slip from seismological data faces additional uncertainties. Seismic slip is inferred from seismological data, subject to the caveats mentioned. Estimation of aseismic slip compounds the uncertainty, because it is done by subtracting the estimated seismic slip from an assumed plate convergence rate, so different plate motion models give different results [*Stein et al.*, 1986]. Moreover, crustal shortening should be subtracted from the assumed convergence before estimating the aseismic slip rate. This correction, which is not done in seismological studies, may be 10-20% of net convergence.

Interpreting seismic slip estimates in terms of the physics of the plate interface has proved equally challenging. It has been suggested that coupling depends on the rate and age of subducting crust (highest when young crust subducts rapidly), trench sediments (least when sediments are thickest), and the normal stress on the plate interface (lowest when stress is low) [*Scholz*, 1990]. Although these ideas seem plausible, efforts to correlate the seismic slip fraction with convergence rate or plate age find no clear pattern [*Pacheco et al.*, 1993], and the sediment and stress arguments have yet to be rigorously tested. Hence at this point "seismic coupling" is a useful kinematic term, but its relation to the physics of the interface remains problematic.

Geodetic data provide another way of investigating these issues because the deformation of the overriding plate (Figure 13) depends on the geometry of the locked plate interface and the rate at which slip accumulates, to be released in future earthquakes [*Savage*, 1983]. The locked slip rate inferred from modeling the data can be used to infer a coupling fraction analogous to the seismic coupling fraction.

When such fractional coupling is reported the value depends not only on the data and the model, but the reference used to define it. For example, across a plate boundary zone with 100 mm/yr net motion, of which 20 mm/yr goes into deforming the upper plate, 60 mm/yr of locked slip at the trench might be quoted as 60% or 75% coupling, and this fraction would change if estimates of the total plate motion changed even if the locked slip estimate remained the same.

Studies for trenches including Indonesia [*Prawirodirdjo et al.*, 1997], Peru-Chile [*Norabuena et al.*, 1998; *Bevis et al.*, 2001], Japan [*Le Pichon et al.*, 1998; *Nishimura et al.*, 2000], Cascadia [*McCaffrey*, 2000; *Miller et al.*, 2001b], Ecuador [*Kellogg et al.*, 1997], Middle America [*Dixon*, 1993; *Lundgren et al.*, 1999; *Lowry et al.*, 2001], and Alaska [*Freymueller et al.*, 2000] show a range of locking, from essentially zero to essentially full. Although surface observations record the integrated effect of possibly-variable properties of the plate interface (partial locking presumably reflect some patches locked and others free-slipping), GPS data in some places are already dense enough that detailed two-dimensional images of the plate interface are being developed [*Freymueller et al.*, 2000].

Such data should significantly improve our understanding of the plate interface locking process. The relation between the locked zone and the seismogenic part of the interface may not be simple, because interface properties, especially down-dip extent, inferred from geodetic data may differ from those inferred from large earthquakes [*Savage*, 1990]. For example, geodetic data show silent slip events, not associated with seismicity, below the seismogenic interface [*Dragert et al.*, 2001]. The difference between seismological and geodetic values and the locked extent of the interface inferred from the different techniques presumably reflects its physical nature, and has implications for the mechanics of large subduction thrust earthquakes. The down-dip extent may reflect a thermally activated process, such that deformation passes from brittle or semi-brittle to ductile or semi-ductile. Alternatively, it may reflect the onset of metamorphic reactions (thermally or pressure-driven) that release water and increase pore pressure sufficiently to reduce brittle interaction (thermally activated creep should ultimately limit seismic interaction). Geodetic data are already playing a major role in constraining thermo-rheological models of the interface, especially in Cascadia where data from large earthquakes are not available [*Hyndman and Wang*, 1993]. New seafloor GPS technology combining acoustic underwater positioning and high quality ship-board GPS positioning [*Speiss et al.*, 1998] is likely to be of great value, because data near the trench can improve the resolution of the shallowest parts of the interface.

In addition, GPS data are starting to provide our best data to date on the full complexity of the temporal and spatial deformation field associated with large earthquakes at plate boundaries. Until recently, although conventional geodesy indicated that earthquakes were followed by complex post- and inter-seismic deformation, the difficulty of making these measurements led to relatively sparse data. GPS data now show significant motion, not detectable seismologically, following earthquakes [*Heki et al.*, 1997, *Bürgmann et al.*, 1997]. The effects can be dramatic: in the Kenai Peninsula, GPS data show that large deformations from the 1964 Prince William Sound area ($M_w = 9.3$) earthquake, the second largest ever recorded, are still occurring [*Cohen et al.*, 1995; *Freymueller et al.*, 2000], and appear to reflect the slip distribution in the earthquake.

Data over time show the history of postseismic motion. For example, GPS data show the motion along the Rivera-North America subduction interface following the October 1995 $M_w = 8$ Jalisco earthquake [*Hutton et al.*, 2001]. Inversion of site displacements to infer the slip distribution during the large earthquake yield results similar to analyses of seismic waveforms. After several years inland sites continue the expected seaward postseismic motion consistent with the earthquake, whereas the coastal sites have essentially stopped moving and should soon resume their expected landward interseismic motion.

Such data offer the prospects of learning a great deal about the deformation process. This task is challenging, as illustrated by the diverse interpretations of the postseismic GPS and InSAR data following the Landers earthquake. These data have been interpreted as reflecting processes including afterslip on the fault [*Shen et al.*, 1994; *Wdowinski et al.*, 1997; *Savage and Svarc*, 1997], fault-zone collapse [*Massonnet et al.*, 1996], poro-elastic rebound [*Peltzer et al.*, 1998], and viscoelastic relaxation of the crust and upper mantle [*Deng et al.*, 1998; *Pollitz et al.*, 2000]. However, as data accumulate and modeling advances, understanding of different possible postseismic effects and the rheology of the lithosphere and asthenosphere should improve. These issues are of broad importance to the earthquake process, given the emerging view that stress transfer between faults [*Stein et al.*, 1994] and stress waves traveling large distances [*Pollitz et al.*, 1998] may contribute to earthquake triggering.

Mountain Building

Mountain building and uplift are spectacular consequences of plate convergence. As shown in Figures 7, 11 and 12, GPS data are giving good views of the horizontal motions associated with mountain building. Such data can be combined with seismological and geological data to model the mountain building process. However, although horizontal motions that produce crustal thickening are important in mountain building [*Isacks*, 1988; *Dewey and Lamb*, 1992; *Hindle et al.*, 2002; *Hindle and Kley*, this volume; *Klosko et al.*, this volume], other effects may also be significant. In the Andes, for example, shortening alone may be insufficient [*Kley and Monaldi*, 1998], and processes including magmatic addition [*Lamb and Hoke*, 1997; *James and Sacks*, 1999], thinning (delamination) of the mantle lithosphere [*Pope and Willett*, 1998], lower crustal flow, [*Kley and Monaldi*, 1998], and deep mantle flow [*Russo and Silver*, 1996] may also contribute. Hence the amount of uplift is likely to differ from the approximately 10-15% of the shortening predicted from simple isostasy.

As a result, topographic and paleotopographic data are increasingly being integrated into these studies. One approach is to use the height and form of mountain ranges to constrain models of the processes that produced them. Topography is not just a passive recorder of these processes due to the feedback mechanisms by which elevated topography causes gravitational spreading stresses [*Molnar and Lyon-Caen*, 1988], erosion [*Chase*, 1992], and regional climatic change [*Ruddiman and Kutzbach*, 1989], which then affect the topography [*Horton*, 1999; *Montgomery et al.*, 2001]. Hence topography is used both as an input to models of the forces involved in mountain building [e.g., *Liu et al.*, 2000; this volume] and as a constraint that models are required to reproduce [*Wdowinski and Bock*, 1994; *Yang et al.*, 2001].

However, the modeling is nonunique not only because of the choice of rheologic and force parameters, but because it constrains the model only to match the present topography. Hence it seems likely that models can be better constrained by requiring that they be consistent with both the shortening and uplift histories, and thus offer insight into time-variable aspects of convergence and mountain building [e.g., *Norabuena et al.*, 1999; *Hindle et al.*, 2002; *Hindle and Kley*, this volume]. Moreover, the time at which mountains reached various fractions of their present elevation is important for modeling their effects on climate [*Ruddiman*, 1997].

Geologic uplift data are illustrated in Figure 15 which shows paleoelevation estimates for the Andes synthesized from techniques including sedimentology, geochemistry, volcanology, paleobotany, geomorphology, and geochrology. The data show that as the locus of shortening and volcanism migrated eastward, uplift also migrated eastward. As indicated, these estimates have large uncertainties. Some are due to uncertainties specific to individual techniques. For example, fission track analysis uses the unroofing rate inferred from the time since the minerals cooled below their annealing temperatures as a proxy for the uplift rate, so the inferred rate depends on specific assumptions [e.g., *Benjamin et al.*, 1987, *Masek et al.*, 1994]. Similarly, inferring paleoaltitude from paleofauna requires both botanical assumptions about the relation of past species to their nearest living present relative, and the assumption that ancient climates are the same as today [*Gregory-Wodzicki*, 2000]. Indicators based on the history of the landscape face the tradeoff between climatic and tectonic effects [*Molnar and England*, 1990; *Burbank and Anderson*, 2001]. Beyond these technique-specific issues, there is the general challenge of distinguishing the uplift of individual points (rock uplift) from that of the regional elevation (surface uplift), which are the same only if no erosion occurs [*England and Molnar*, 1990; *Burbank and Anderson*, 2001].

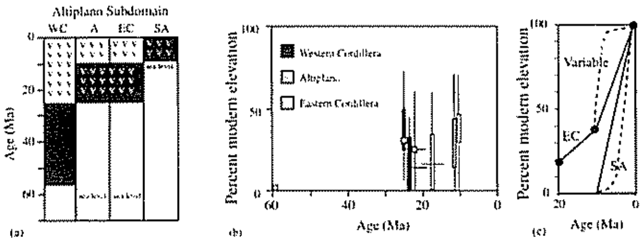

Figure 15. a) General timing of compression (black), arc magmatism ("v"'s), and last marine facies ("sea level") for the Western Cordillera (WC), Altiplano (A), Eastern Cordillera (EC), and SubAndean belt (SA) [*Gregory-Wodzicki*, 2000]. b) Paleoelevation estimates and tectonic events [*Gregory-Wodzicki*, 2000]. c) Schematic illustration of possible uplift histories constrained by average paleoelevations (dots) for Eastern Cordillera and SubAndean belt, showing that actual rates (dashed) could have significantly exceeded their long-term averages (solid).

An interesting possibility is that the uplift rates that would be inferred from the difference between paleoelevations at different times may be minimum values for two reasons. First, because of erosion, even if the actual uplift rate were uniform it would exceed that inferred from the difference in paleoelevations. Second, if the locus of shortening and uplift has been migrating, the average rate inferred would likely be an average over a period of relatively rapid uplift and a period of slower (or no) uplift. However, the actual rate of uplift may be the most crucial rate for understanding its dynamics.

Hence an exciting prospect is using space geodetic data to directly measure present uplift rates. To date, such measurements have focused on the horizontal for two reasons. First, in most tectonic environments horizontal crustal motions are significantly faster than vertical motions. Second,

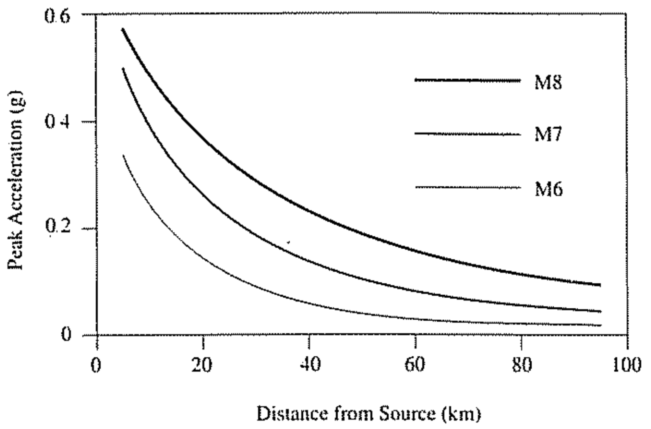

Figure 16. Comparison of the predicted strong ground motion as a function of distance from magnitude 6, 7 and 8 earthquakes in the western United States. A smaller nearby earthquake can give shaking greater than that from a larger one further away. 0.2 g corresponds approximately to the onset of significant building damage, depending on factors including construction type and site conditions [*Tsai et al.*, 2001]. The curves are computed from models by *Sadigh et al.* [1997] and modified after *Stein and Wysession* [2002].

the precision of horizontal measurements is about three times that of vertical ones, where poor modeling of parameters such as tropospheric water vapor, atmospheric and oceanic loading, etc. have a more pronounced effect. Hence although some vertical measurements are available near subduction zones, vertical motions in mountain ranges are rarely measured except where they can be derived from conventional geodesy using precise leveling [*Jackson and Bilham*, 1994]. Given the effort and difficulties involved, such measurements are rare. However, the increasing precision of vertical velocities derived from continuous GPS measurements will likely soon yield estimates precise enough for tectonic studies.

PLATE BOUNDARY ZONES AND SOCIETY

The recognition that plate boundaries are broad zones has implications for geologic hazards to society. Analysis of United Nations Population Division data for 2001 (World Urbanization Prospects: 2001 Revision) indicates that about 40% of Earth's population lives within the boundary zones shown in Figure 1. They are thus vulnerable to hazards not just at the nominal boundary, but over the broad boundary zone. This effect has been recognized for some time in dealing with subduction zone volcanism, which occurs at distances up to a few hundred kilometers landward of the trench [*Hatherton and Dickinson*, 1969]. In recent years it has been increasingly recognized that seismic hazards also extend across the boundary zone. Because ground shaking decays rapidly with distance (Figure 16), nearby smaller earthquakes within a boundary zone can be more damaging than larger but more distant ones on the nominal boundary. Hence the Los Angeles area is vulnerable to both nearby earthquakes like the 1994 Northridge (M_w = 6.7) or 1971 San Fernando (M_s = 6.6) earthquakes and larger ones on the more distant San Andreas Fault, such as a recurrence of the 1857 Fort Tejon earthquake which is estimated to have had M_w about 8. Similarly, the earthquake hazard in the Seattle area involves both great earthquakes at the subduction interface and smaller, but closer, earthquakes in the subducting Juan de Fuca plate (like the 2001 M_w = 6.7 Nisqually earthquake) or at shallow depth in the North American plate. As a result, studies of the kinematics and dynamics of plate boundary zones will also contribute to understanding the geologic hazards there.

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G. F. Sella, S. Stein, Department of Geological Sciences, Northwestern University, 1847 Sheridan Road, Evanston, IL 60208, USA (sella@earth.northwestern.edu, seth@earth.northwestern.edu)