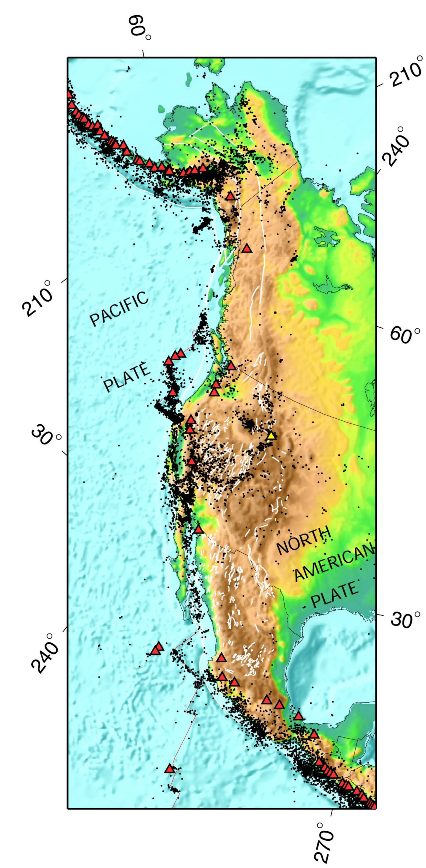
THE PLATE BOUNDARY OBSERVATORY

Creating a Four-Dimensional Image of the Deformation of Western North America

A white paper providing the scientific rationale and deployment strategy for a Plate Boundary Observatory, presented by the PBO Steering Committee to the National Science Foundation. Based on input from the PBO Workshop held October 3-5, 1999.



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Executive Summary

The Earth's tectonic plates are in continual motion, creating a constantly changing mosaic as the plates move away from, past, or beneath each other at plate boundaries. While these plate-boundary zones are razor-thin in classical plate tectonic theory, it is clear that our own continent's boundary, between the Pacific and North American plates, is not. These two plates, like the two hands of a mighty sculptor, have produced the rugged landscape of western North America: the Rocky Mountains, the Sierra Nevada, the Basin and Range province, the majestic volcanic peaks of the Cascades and Aleutians, and the San Andreas Fault system. These plate-boundary forces are most vividly manifested by the violent shaking of earthquakes, capable of leveling cities in a matter of seconds, and volcanic eruptions that spew forth molten rock from the Earth's hot interior. The questions that we ask of these plate-boundary zones are as old as humanity itself. How are mountains formed, and how do they evolve? How do earthquakes occur? How do volcanic eruptions occur? What forces drive these often-catastrophic changes?

The theory of plate tectonics has provided an important framework and partial answers to these questions. We know that earthquakes are due to slip on faults produced by the relative motions of the plates, and that volcanism is the result of three basic processes: subduction, rifting, and hot spot activity. We also know that there are a variety of forces capable of driving the deformation that builds mountains. Yet this theory has left some of the most important questions unanswered. What is primarily missing is the dimension of time. What determines the sequencing of seismic events? When and where will the next volcano erupt? What controls the rate and spatial pattern of plate-boundary deformation, and how does it evolve over time?

Much as a sculptor learns his art by watching a master in the act of creation, we as Earth scientists can make dramatic progress toward answering these basic scientific questions by improving our ability to observe the constantly changing, four-dimensional image of western North America. We seek to monitor the motions of mountains as they form and collapse, map the strain accumulation that leads to a major earthquake, and track the movement of magma that ascends from depth and then breaches the surface in a volcanic eruption.

In order to understand these plate boundary processes, it is thus essential to observe plate-boundary deformation over a very broad range of temporal and spatial scales. We thus propose to deploy a **Plate Boundary Observatory (PBO)**, a facility that would create this four-dimensional image in sharp and unprecedented detail.

The range of spatial scales to be examined is vast: from meter-sized fault planes of small earthquakes to the thousands of kilometers of large fault systems and mountain chains. The range of timescales of plate-boundary deformation is even larger, from the several-second duration of a devastating earthquake to the millions of years necessary to build mountains. These spatial and temporal scales embrace much of Earth science, from seismology at short periods, to geodesy at longer periods and larger spatial scales, to geology at still greater timescales. While the seismological end of the spectrum will be well-covered by regional seismic networks, the new Advanced National Seismic System, and the proposed **USArray** component of the **EarthScope** initiative, there are large gaps at the geodetic and geologic end of the spectrum that must be filled if our understanding of plate-boundary dynamics is to be substantially advanced. Thus, the core of PBO is a geodetic observatory, with an extension to periods longer than several decades by means of geological techniques, and to the subsurface by means of the imaging capabilities of seismology.

The timing of this initiative is enabled by the fortuitously coincident culmination of progress in several areas: the development of complementary geodetic instruments that together are capable of spanning the required broad temporal range of plate boundary deformation, and advances in geologic measuring capability that provide strain data on the thousand-to-million-year time scale. PBO involves an integration of these new technologies, not heretofore possible, into an Earth observing system unprecedented in scope and resolution, qualities that traditionally in science have led to new insight and the discovery of new phenomena.

The core instrumentation request is for a geodetic observatory consisting of a carefully designed and integrated network of strainmeters and GPS receivers. Taken together these two instrument types span the expected broad temporal spectrum of plate boundary deformation. The strainmeters are ideal for recovering short-term transient deformation, from minutes to a month, and will consequently play a central role in observing phenomena that accompany and precede earthquakes and volcanic eruptions. GPS is ideal for time scales greater than a month, thus covering long-period transients, such as those associated with viscoelastic relaxation following an earthquake, as well as decadal estimates of strain accumulation and plate motion and their spatial variations. Only an integrated deployment of these two instrument types is capable of providing temporal resolution over the full set of timescales from minutes to decades, at the necessary spatial resolution and areal coverage of the plate-boundary system.

We propose a two-tiered deployment strategy of this instrumentation:

- A backbone of sparsely distributed (~100-200 km spacing) continuous GPS receivers to provide a synoptic view of the entire North American plate boundary deformation zone.

- Clusters of instruments to be deployed in areas that require greater spatial or temporal resolution, such as the major fault systems and magmatic centers.

The full PBO instrumentation network would consist of 1,275 continuously recording GPS receivers and 245 strainmeters, including the 400 existing GPS and 45 existing strain instruments. 100 new GPS receivers would be allocated to the backbone, and the rest of the instrumentation would be assigned to clusters. Specified targets for clusters are the San Andreas Fault system, and six magmatic centers, including Yellowstone, Long Valley, and two Alaskan and two Cascade volcanoes. 400 GPS receivers and 175 strainmeters would be deployed along the San Andreas Fault system. Each of the volcanic centers would, on average, involve a deployment of 15-20 GPS receivers and 4 strainmeters (100 GPS instruments and 25 strainmeters total). 275 GPS receivers (or a mix of GPS and strainmeters) would be available for other targets, the choice of which would be decided competitively on the basis of scientific merit. Finally, we would also need 100 additional GPS receivers for use in survey-mode, for densifying areas not sufficiently covered by continuous GPS, and for rapid response capability following an earthquake or volcanic eruption.

The problems that we seek to address require more than geodetic data. Earthquakes and volcanic eruptions originate in the crust, and we want to measure deformation in the nucleation zone of earthquakes and where magma is forming and migrating. Seismology is our best tool for studying the depth-dimension of deformation. Microearthquake activity is a proxy for aseismic deformation in the seismogenic zone, and for this reason, borehole seismometers are included with the strainmeters. Seismic tomography can provide images of subsurface deformational structures, such as faults. USArray will provide much of the capability for this imaging. We also need to know the history of deformation over thousands to millions of years, well beyond the timescales available through geodesy. Geological measurement provides this crucial context for several problems. For example, paleoseismology yields a history of earthquake activity throughout several earthquake cycles, information necessary both for understanding the earthquake process and for accurately assessing earthquake hazard. Comparison of decadal and Holocene patterns of deformation allows us to study how the plate-boundary zone evolves over time. A critical requirement for measuring this long-term deformation is the acquisition high-resolution topographic data, through airborne imaging. Based on the topographic data, we propose to systematically map fault offsets and other deformational structures in regions either presently active or active in the Holocene. Such data will provide the basis for the most comprehensive and detailed estimates of Holocene strain ever obtained.

With the resulting four-dimensional image of plate boundary deformation, we will be poised to answer the following scientific questions:

- What are the forces that drive plate-boundary deformation?
- What determines the spatial distribution of plate-boundary deformation?
- How has plate-boundary deformation evolved?
- What controls the space-time pattern of earthquake occurrence?
- How do earthquakes nucleate?
- What are the dynamics of magma rise, intrusion, and eruption?
- How can we reduce the hazards of earthquakes and volcanic eruptions?

Finally, PBO will be an agent of discovery. This Earth telescope, focused on the continually changing face of our North American continent, will reveal features and processes never before seen. We are ready to begin the exciting journey toward a new understanding of our dynamic landscape.

Preface

This document is based on the outcome of the PBO workshop, held October 3-5, 1999. It has been jointly written by the PBO Steering Committee, whose members are

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1.0 Introduction

The development of the theory of plate tectonics in the 1960's was the seminal breakthrough in the geosciences in the twentieth century. The theory holds that the outermost 100 km of the solid Earth is composed of a relatively small number of internally rigid plates (lithosphere) that move over a weak substrate (asthenosphere). Relative motions between the plates are 1 to 10 centimeters per year. The basis of these plate velocities is the pattern of linear magnetic stripes in the oceanic basins, generated at divergent plate boundaries, and the history of the reversals in polarity of the Earth's dipole magnetic field responsible for those anomalies.

The precision of plate kinematics obtained with this theory was dramatically confirmed in the 1990s with the advent of space geodetic techniques. The comparison showed that most interplate velocities obtained geodetically were in agreement with those inferred from seafloor magnetic anomalies to within a few percent, an astonishing result given that the geodetic velocities are averages over ~10 to 15 years, while the plate tectonic velocities are averages spanning one to several million years, a difference of five orders of magnitude in timescale.

A basic tenet of the plate theory is that most earth deformation is concentrated within narrow zones at plate boundaries. However, it is now clear that plate boundary deformation zones, especially those involving continental lithosphere, are often quite broad – extending thousands of kilometers into plate interiors, such as along the Alpine-Himalayan belt of Europe and Asia and the western mountain chains of North and South America. Nearly all present-day tectonic activity and most non-meteorological natural hazards, including earthquakes and volcanic eruptions, are concentrated within these zones. For example, the active boundary zone between the Pacific and North American plates in the western United States covers one-third of the North American continent (western cordillera) and includes such diverse tectonic and volcanic elements as the Rocky Mountains, Basin and Range province, Yellowstone hotspot, Cascade and Aleutian volcanic arcs, Sierra Nevada, and San Andreas Fault system (**Fig. 1**).

It is also clear that the inexorable and quasi-steady motions of the tectonic plates, while providing an important kinematic boundary condition, provide little insight into a broad variety of plate boundary phenomena, especially the episodic behavior of earthquakes and volcanoes and the complex patterns of lithospheric deformation. For example, plate tectonics does not specify the forces that drive plate motion, or the way in which stress and strain are transferred through the plates. It does not explain how steady plate motion ultimately produces time-dependent phenomena, most notably the triggering of individual earthquakes and volcanic eruptions. Nor does plate tectonic theory account for the locations of faults or magmatic centers. Steady plate motion, in fact, leads to spatially variable and often transient styles of deformation. This complex deformational system reflects spatial and temporal variations in geologic structures, regional stress, and crust/upper mantle rheology. An observatory, capable of providing a comprehensive four-dimensional image of this system, holds the key to addressing problems that are both fundamental and of direct concern to society.

Imagine an Earth observatory that watches the North American continent deform, detects the growth and collapse of mountains, monitors active faults as they accumulate and release strain during major earthquake cycles, and searches for the tell-tale signs of magma welling up from depth in preparation for a volcanic eruption. Imagine, further, peering beneath Earth's surface into zones where earthquakes and volcanic eruptions actually originate, and back in time to trace the changing face of the continent through the sweep of geologic history. The potential for scientific

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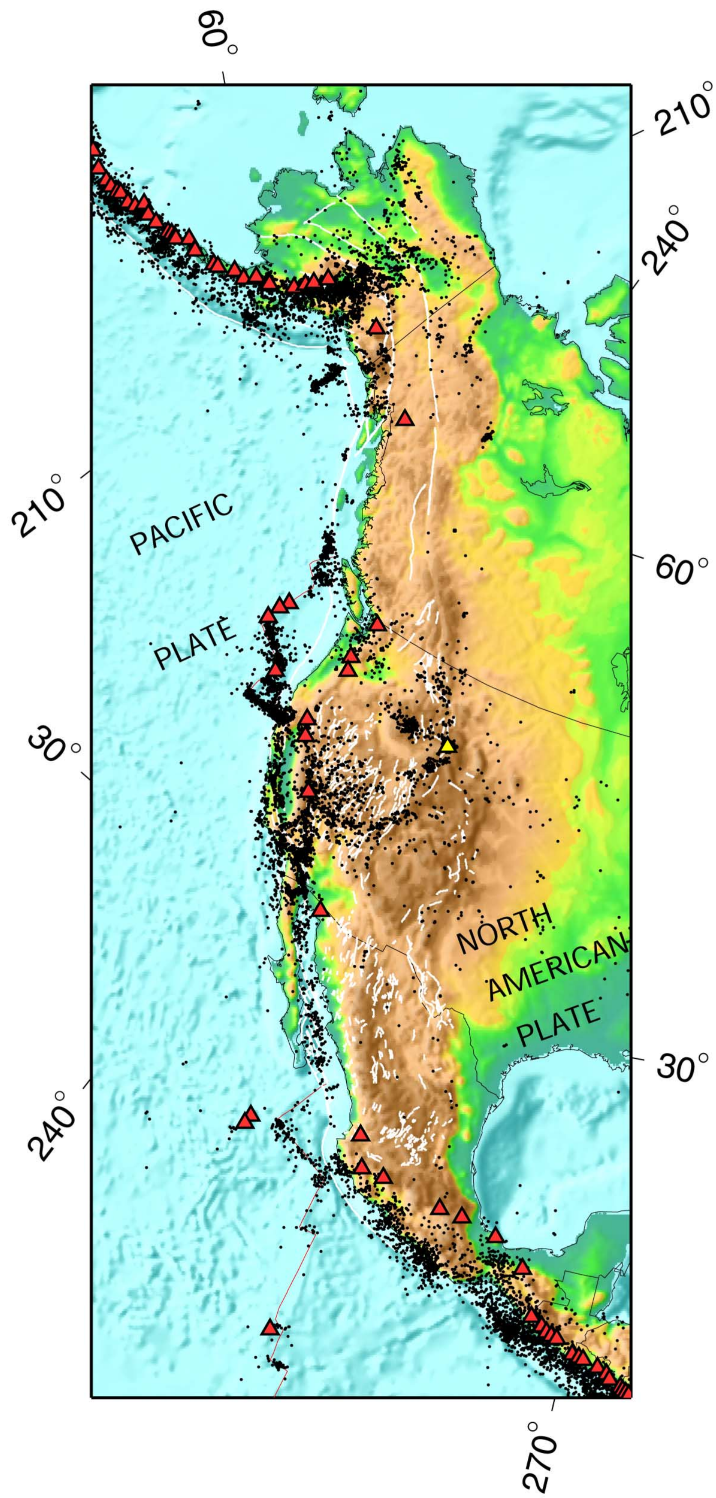


Figure 1: Shaded relief map of western North America with active faults (white lines, Peter Bird, personal communication), recent seismicity (last 10 years, black dots), and active volcanoes (those with recent eruptions or signs of tectonic activity, red triangles; yellow triangle is Yellowstone). The San Andreas Fault system is located at the highly seismogenic western edge of the plate boundary zone.

discovery and for advancing our knowledge of continental dynamics would be immeasurable. Such a facility would lead not only to a major increase in our general understanding of plate boundary processes, but also to dramatic progress in understanding the occurrence of earthquakes and volcanic eruptions, with the possibility of ultimately forecasting their occurrence.

We propose to deploy such an observatory, called the Plate Boundary Observatory (PBO). PBO is an integrated, multi-component instrumental array that will provide a fully four-dimensional representation of the actively deforming plate boundary zone of North America **(Fig. 1)**. The range of spatial scales to be examined is large, from meters, characteristic of the smallest earthquakes, to hundreds of kilometers, as represented by plate-bounding structures such as the San Andreas Fault system or Cascade and Alaska/Aleutian subduction zones, to the thousands of kilometers dimension of the full plate boundary zone. The range of timescales over which plate boundary deformation occurs is even greater, from the seconds it takes for a devastating earthquake, to the millions of years necessary to build mountains. These spatial and temporal scales, and the active tectonic or deformational processes they represent, also span much of Earth science, from seismology at short periods, to geodesy and geology at longer periods and larger spatial scales. While the seismological end of the spectrum will be well-covered by regional seismic networks, the new Advanced National Seismic System, and the proposed USArray component of the EarthScope initiative, large gaps exist at the geodetic and geologic end of the spectrum. Thus, the core of PBO will be a geodetic observatory, with an extension to periods longer than about 100 years using geological techniques, and to depth, using the imaging capabilities of seismology and geodetic inversion.

This initiative is timely because of breathtaking technical advances over the last decade in the disparate areas of precise GPS positioning, strain measurement, topographic imaging, and radiometric dating relevant to active tectonic phenomena, techniques which have heretofore been applied only in uncoordinated, piecemeal fashion across the plate boundary. PBO will represent the crystallization of these new technologies into an unprecedented 'Earth telescope' that will reveal phenomena and processes that we have yet to even dream about. PBO will provide crucial information complementary to the Phase I components of EarthScope, USArray and the San Andreas Fault Observatory at Depth (SAFOD).

2.0 Scientific Goals of PBO and Required Observations

Three basic areas of research will benefit dramatically from the deployment of the PBO. They are:

- Understanding how the Earth's tectonic plates interact.
- Understanding the physical basis for earthquakes.
- Understanding the magmatic processes that lead to volcanic eruptions.

While these three areas are traditionally studied separately, they clearly have strong interactions. The recent evolution of deformation is closely linked to and largely defined by the migration of earthquake activity. The study of postseismic deformation provides perhaps the best constraint on crust and upper mantle rheology, and hence is crucial to studying the underlying forces driving earthquakes and plate boundary deformation in general. The occurrence of earthquakes and volcanic eruptions may be causally linked in some cases. These phenomena also represent processes on a broad range of scales, from the entire plate boundary zone to the nucleation zone of an earthquake. Dramatic progress will come through the deployment of an observing system that is unprecedented in its coverage of an entire plate boundary, as well as its spatial and temporal resolution. Here we discuss these three major scientific areas, provide a brief discussion of our current understanding and its limitations, and demonstrate how the PBO will dramatically advance that understanding.

Each of the general scientific themes requires measurements of the plate boundary deformation field at different spatial and temporal scales. Thus the PBO must, at a minimum, be capable of providing the necessary observations for addressing these problems. There are several measurement techniques that must be integrated. The surface strain field would be measured directly with geodetic instrumentation. Instrumentation must provide enough coverage of the plate boundary zone to capture its behavior at the scale of the entire system, and of sufficient station density for detecting localized (e. g., fault- and volcano-specific) phenomena. It must also possess the necessary temporal resolution to detect deformation transients, ranging from those associated with fast and slow earthquakes and magmatic events, to interseismic strain accumulation and postseismic relaxation. For studying large-scale processes, such as mountain building, it is probably sufficient to examine spatial variations in the geodetically-determined decadal strain rate. Geological measurements of deformation must be made to place the geodetic observations into a longer-term context. Measurements of this long-term deformation (beyond 100 years) dramatically extend the observable timescale, which is valuable to all of the areas of plate boundary research, from extending earthquake catalogs, to characterizing the long-term and Quaternary evolution of plate boundary zones through comparison with geodesy. Because both earthquakes and magmatic processes originate beneath the surface in the crust, it is crucial to extend the surface deformation measurements to depth. While we cannot measure strain at depth directly, surface deformation measurements can be used to invert for seismic and volcanic sources at depth. In addition, we can measure proxies for strain, such as microearthquake activity, that provide detailed spatial information on the distribution of strain at depth.

2.1 Plate Boundary Dynamics and Evolution: Establishing a Framework

Forces driving deformation

While plate tectonics does not adequately explain plate boundary deformation, it does provide a useful framework for studying it. The remarkable steadiness of relative plate motion, even on timescales as short as a decade, represents a fundamental kinematic boundary condition for the study of plate boundary deformation. This constant motion has produced a complex pattern of continental deformation. One of the basic goals of the PBO is to determine the forces responsible for this deformation and to assess their relative contributions. The relevant forces include plate boundary forces applied at the edge of plates, forces associated with coupling to the mantle flow field below (such as basal drag or driving shear stresses), and gravitational forces due to lateral density variations. We thus ask:

- *What are the forces that drive distributed plate boundary deformation, namely, the relative contributions of plate boundary forces, buoyancy forces, and basal forces?*

Role of rheology

While most plate tectonic forces act on the scale of an entire plate boundary, plate boundary deformation structures display a variety of spatial scales, ranging from the entire plate boundary zone to the offsets on individual faults. This variability is due in large part to spatial differences in rheology, the quantity that defines the relationship between stress and strain in the crust and mantle. It is thus critical to determine the rheology in order to relate the observed deformation to the driving forces, and to explain the particular structural characteristics of western North America. Rheology is a sensitive function of rock type and temperature and may have many possible forms in the plate boundary zone. There has been extensive experimental research aimed at constraining the rheology of crustal and mantle materials. Yet, these experiments are, of necessity, performed at temporal and spatial scales that are orders of magnitude smaller than those relevant to plate boundary processes. Making progress in determining the appropriate rheology thus requires relating rock deformation on laboratory scales to deformation on scales that characterize large scale patterns of regional tectonic deformation. We thus ask:

- *What is the spatial distribution of plate boundary deformation rate and how is it related to spatial variations in stress and rheology?*

Time dependence

If Pacific/North American relative plate motion has been constant over the last several million years, does that mean that the corresponding strain-rate field within the plate boundary zone is also in steady state? Is there evolution over thousands of years? Over millions of years? Does tectonic activity migrate? Are there deformation transients? If so, what are the characteristic timescales? We know that there are short-term transients, in the form of earthquakes and related deformation that we can measure directly from geodesy. We will focus on this important topic in the next section. But is there temporal variability at longer timescales (greater than 100 years) that average over several earthquake cycles? Such variability would be important to verify, since the time dependence would require either a change in the force balance (e.g., by a changing the gravitational force due to mountain building/collapse), the property of a time dependent rheology (e.g., viscoelastic relaxation), or a temporal change in the rheology itself (say through a magmatically induced change in crust/mantle temperature). We thus ask:

- *Is there long-term transient deformation within the plate boundary zone, and if so, what are the characteristic temporal scales and underlying causes?*

Current constraints and limitations

The existing but very limited data allow us to begin to address these basic questions and also to highlight where new observations are needed to make significant progress. Shown in **Fig. 2a** is a velocity field across a portion of the North American-Pacific boundary zone, derived by combining GPS and VLBI data from various sources (Bennett et al., 1999). Net motion across the zone is essentially that predicted by the global plate motion model NUVEL-1A, which provides a good estimate of plate motion averaged over millions of years. This similarity thus illustrates the constancy of the plate motion driving the deformation. The site velocities reveal several characteristics of the deformation field. The strongest deformation is associated with the strike-slip behavior along the San Andreas Fault system. There are also marked spatial variations in site velocity within the Basin and Range Province. Some of this variation essentially defines the present-day boundary between stable and deforming North America. Northern and southern profiles of the geodetic deformation **(Fig. 2b)** are very different. South of 36ºN, the San Andreas Fault system accommodates most of the plate motion, and little deformation occurs in the Basin and Range, whereas to the north strike-slip motion changes into extension over a broad transition zone. It thus appears that the Intermountain seismic belt **(Fig. 2a)** defines the eastern boundary of the deforming plate boundary zone.

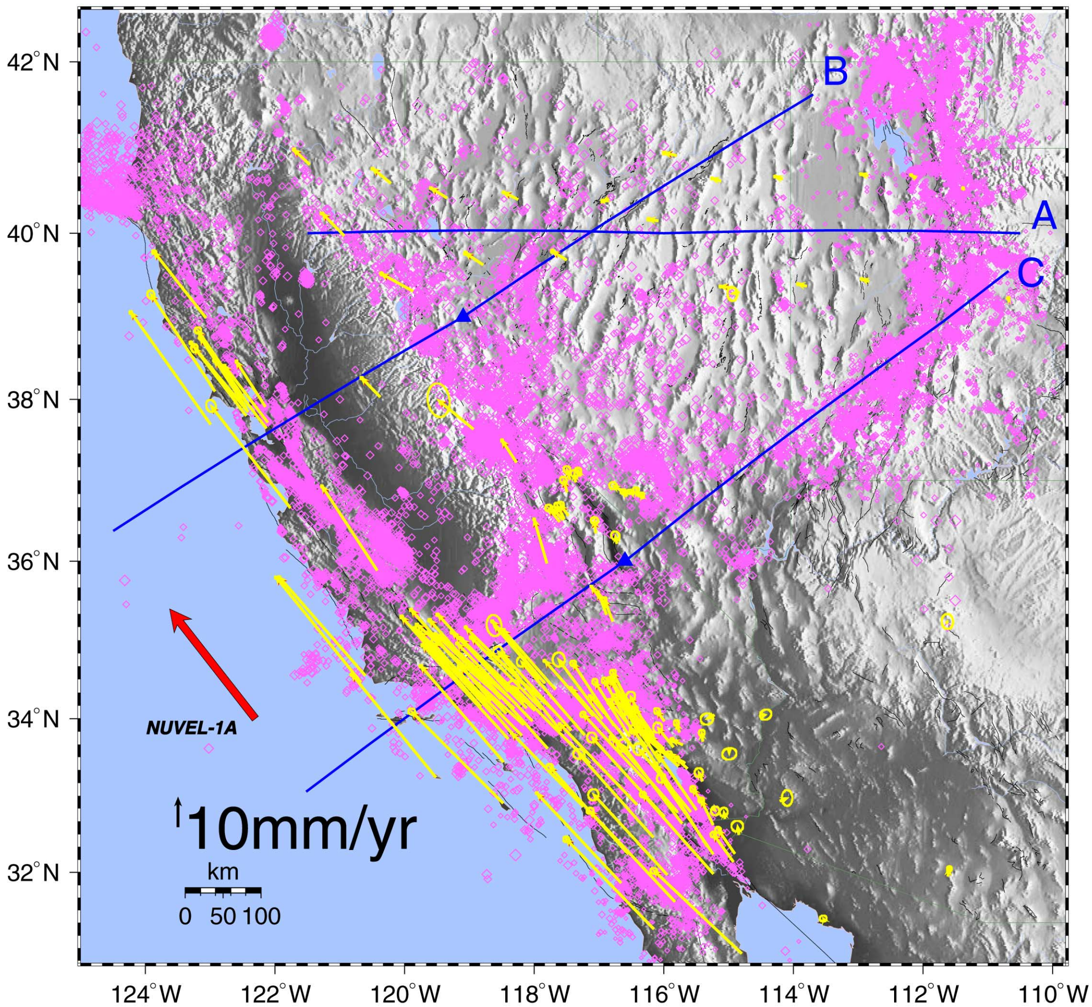
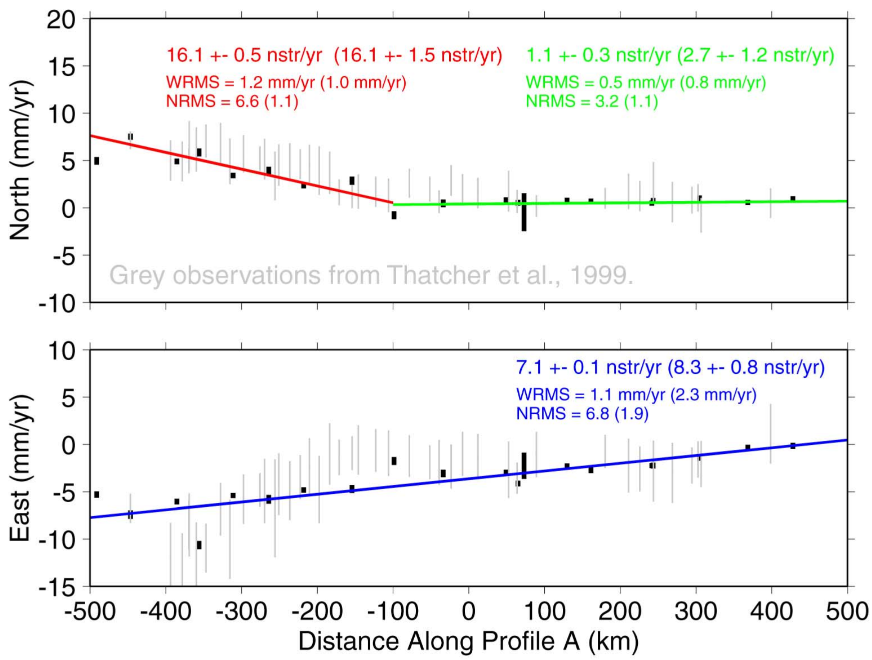


Figure 2a: Velocity field (yellow arrows) across a portion of the North American-Pacific plate boundary zone, derived by combining GPS and VLBI data from a variety of sources. Also shown, seismic activity (pink). Site motions are relative to stable North America (Bennett et al., 1999).



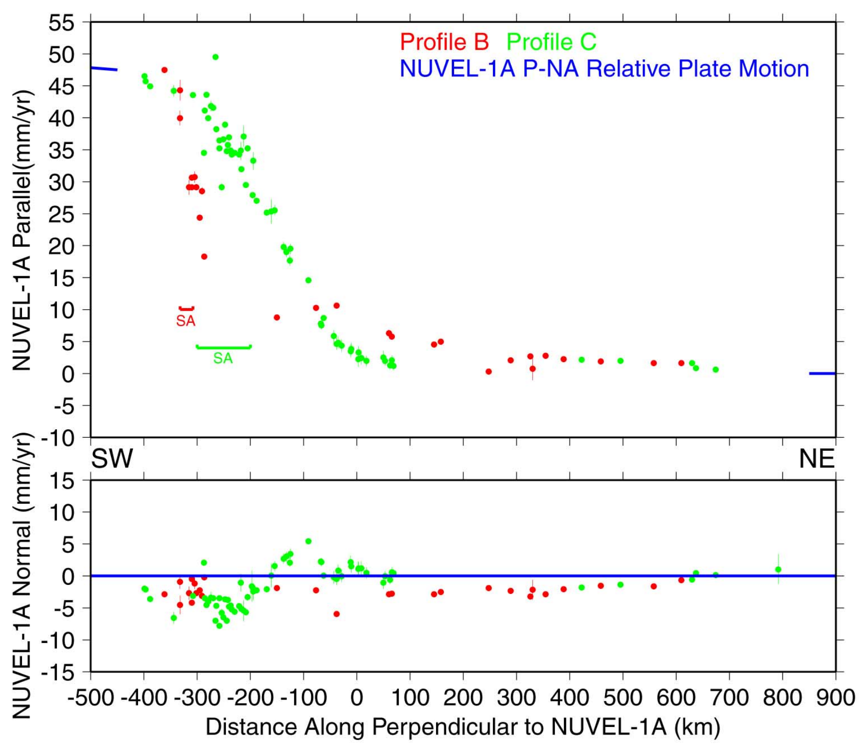


Figure 2b: Site velocity profiles through the data in Fig. 2a. For profiles B and C, data are shown in directions parallel and normal to the expected North American-Pacific relative plate velocity direction, based on the global plate motion model NUVEL-1A (red arrow in Fig. 2a). Negative values along NUVEL-normal profile indicate contraction between the Great Basin and the eastern California shear zone. For profile A, continuously recording GPS data are also compared with survey-mode GPS data of Thatcher et al. (1999) showing excellent agreement.

More generally, these spatial variations provide a means of constraining the dynamics and rheology of the boundary zone. Specifically, the relative importance of topographically-derived buoyancy forces and plate boundary forces can be assessed through modeling the deformation. In such modeling (based on the thin viscous sheet approximation of England and McKenzie, 1982), the observed site velocities are used to compute a strain-rate tensor field which, together with an estimate of the gravitational potential energy from topography and geoid data (Shen-Tu et al., 1998), yield estimates of the relative contributions of these two forces and the spatial distribution of rheology. In a recent example of this approach using available geodetic and geological data for western North America, Flesch et al. (2000) have shown that the deformation in the Basin and Range requires contributions from both gravitational and plate boundary forces. Because the gravitational contribution is approximately known, it is then possible to determine the magnitude of the modeled stresses and, used together with the magnitudes of observed strain-rates, to provide an estimate of effective viscosity structure for the area **(Fig. 3)**. The resulting viscosity field shows significant variability, changing by three orders of magnitude over the region, including low values along the San Andreas Fault and in the eastern California shear zone. The apparent focusing of deformation near the western edge of the Basin and Range, as suggested by Quaternary fault patterns (Wallace, 1984; Dokka and Travis, 1990) and seismicity (Eddington et al., 1987), and confirmed by space geodesy (Dixon et al., 1995; Bennett et al., 1999; Thatcher et al., 1999; Dixon et al., 2000), is suggestive of dramatic rheological differences across the Basin and Range, since a uniform-rheology model does not reproduce this distinctive pattern. There is consequently a very close correspondence between the observed regions of localized deformation and the inferred distribution of effective viscosity.

Dixon et al. (2000) demonstrate a close correlation between maximum surface velocity gradient and maximum gradient in surface heat flow in the western Basin and Range. The correlation suggests that strain is accommodated in weaker regions and indicates a close correspondence between upper crustal weak zones (faults) and regions of hot, weak lower crust and upper mantle. The critical region for both high velocity gradient and horizontal changes in surface heat flow is less than 50 km in width. This result underscores the importance of obtaining a high resolution map of the current velocity field and the ability to make detailed comparisons with other geophysical parameters in order to understand the dynamics of the plate boundary zone. While this kind of modeling constitutes a useful starting point for understanding plate boundary deformation, it also highlights several observational limitations that can only be overcome by the PBO. First, the constraints on the deformation field are patchy, only sampling a small fraction of the plate boundary zone, whereas most of the forces act over scales of hundreds to thousands of kilometers. Broad geodetic coverage would not only place much stronger constraints on the forces involved, but would also enable comparable estimates of rheology over the entire plate boundary region. Second, this formulation ignores possible contributions from basal shear stresses. These stresses are due to the coupling of the plate to the mantle below, and thus could be a significant source of stress. Third, the model constitutes a vertical average of lithospheric stress and rheology. It is only capable of recovering spatial variations in the dynamics over length scales a few times the lithospheric thickness (100 km). This approach thus requires an accurate long-wavelength deformation field. It is consequently important that the observed field is not aliased by inadequately sampled short-wavelength deformation. This problem can be solved by the dense sampling of the deformation field in areas that are suspected of concentrating strain. Moving beyond vertically averaged rheology to resolving its vertical variations requires additional information. One approach is by observing the postseismic deformation field, as discussed below. Since the stress change produced by large earthquakes can be readily inferred seismologically, the spatial and temporal characteristics of post-seismic deformation can be used to constrain such vertical variations in the crust and upper mantle.

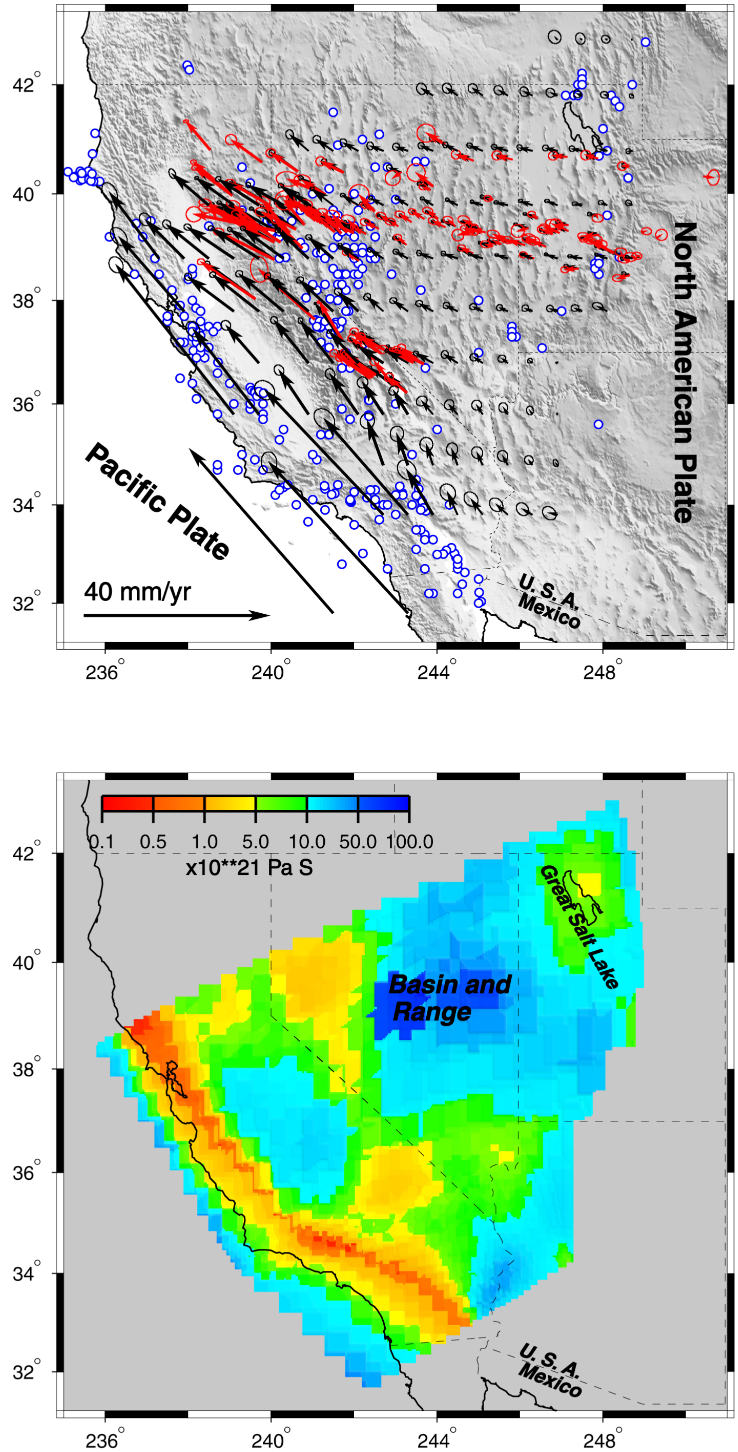


Figure 3: (Top) Self-consistent velocity field (black arrows) from interpolation of GPS and VLBI data. Ellipses represent a 95% confidence limit. Blue circles represent earthquakes recorded from 1850-1998. (Shen-Tu et al., 1998). (Bottom) Vertically averaged effective viscosity determined by dividing the magnitude of the deviatoric stress tensor by the magnitude of the strain rate tensor. The viscosity varies by three orders of magnitude over the region. Low values are seen along the San Andreas Fault and in the California shear zone (Flesch et al., 2000).

What constraints can be placed on long-term transient behavior? Observationally, the most straightforward way to approach this problem is by the direct comparison of the geodetic measurements to be made by PBO with faulting and deformation mapped geologically. Whether or not strain rates are constant when averaged over many earthquake cycles has important implications for active tectonics and the mechanics of strain accumulation, as well as assessing current earthquake hazard.

Limited comparisons between geodesy and Holocene geology suggest a fascinating range of behaviors. On the San Andreas Fault system, geodetic slip-rate estimates are in good agreement with results obtained from paleoseismic studies of earthquake histories and determinations of late Holocene slip rates, suggesting that the San Andreas is in steady state. Within the Great Basin, however, there are indications of temporal variations. Across the central Nevada seismic zone, westernmost Nevada, and the Wasatch Fault system, for example, geodetic estimates of extension rates exceed those based on geological estimates by a factor of three or more. Conversely, some Great Basin faults that are currently not accumulating strain show geomorphic and geologic evidence of significant late Quaternary activity. These data raise the intriguing possibility that tectonic activity migrates on timescales of thousands or tens of thousands of years. If such transient behavior is indeed real, it implies a time-dependent process involving the long-term redistribution of stress and strain.

Precise observation of the present-day strain-rate field over zones that are presently active or have been active over the last few million years is a first-order requirement for studying Holocene or Quaternary time-dependent deformation. This detailed deformation field can then be compared and contrasted to the strain and strain-rate fields that can be obtained from geodesy. There are several techniques for obtaining this ’paleo-strain’ field. The most successful in the 100 to 100,000 year interval is paleoseismology, or the detection and dating of slip on active faults. For example, where fault zones are covered by accumulating sediment, earlier earthquakes disrupt older layers, which are then overlain by undisturbed layers. The age of the earthquake is constrained by dating the youngest disturbed layers and the oldest undisturbed layers. These data can also be “fed back” into models for the geodetic data, incorporating realistic rheology, allowing more accurate estimation of present day fault slip rates, which may be influenced by past earthquake activity and the viscoelastic response of the lower crust. Deformation rates, including slip rates of active faults and shortening rates across active fold belts, and eruptive fluxes of magma can be measured using aerial photography. Tectonic geomorphology provides a means for measuring the often significant strain between major fault zones, or strain not accommodated by surface-breaking ruptures, particularly in areas of horizontal shortening adjacent to the San Andreas transform and Cascadia subduction zone. Using Pre-Quaternary geologic features such as Late Cenozoic basins and large-scale geologic markers (e.g., isopachs and thrust faults), it is possible along much of the plate boundary to reconstruct the strain history on the 1-10 Myr timescale, with increments of strain resolved in 1-4 Myr intervals. This approach has been applied to the central part of the Basin and Range province **(Fig. 4)**, and yields the Late Cenozoic motion of the Sierra-Great Valley block, a large, relatively undeformed crustal fragment along the plate boundary (Wernicke and Snow, 1998).

In summary, a comprehensive, geodetically determined map of deformation-rate, combined with geologic estimates of long-term deformation, will dramatically improve our understanding of driving forces and rheology of the plate boundary zone, and will enable us to clearly document and ultimately understand long-term transient behavior.

Required Observations

- *Detailed mapping of the present day strain-rate field over the entire plate boundary zone in order to constrain the dynamics of continental deformation and the rheology of the lithosphere. This will require a backbone array of continuous GPS stations at 100-200 km spacing, grading into denser arrays across active fault zones.*

- *Detailed characterization of post-seismic deformation of large earthquakes (or other known sources of stress) to obtain meaningful constraints on vertical variations in the rheology of the plate boundary zone.*

- *Geologic estimates of long-term (Holocene and Quaternary) strain-rates. These include, but are not limited to, estimates of fault slip rate across many fault systems to assess the temporal variability of fault activity. Corresponding geodetic estimates of present-day surface deformation are needed in areas that are active today or have been active in the past, based on geologic estimates.*

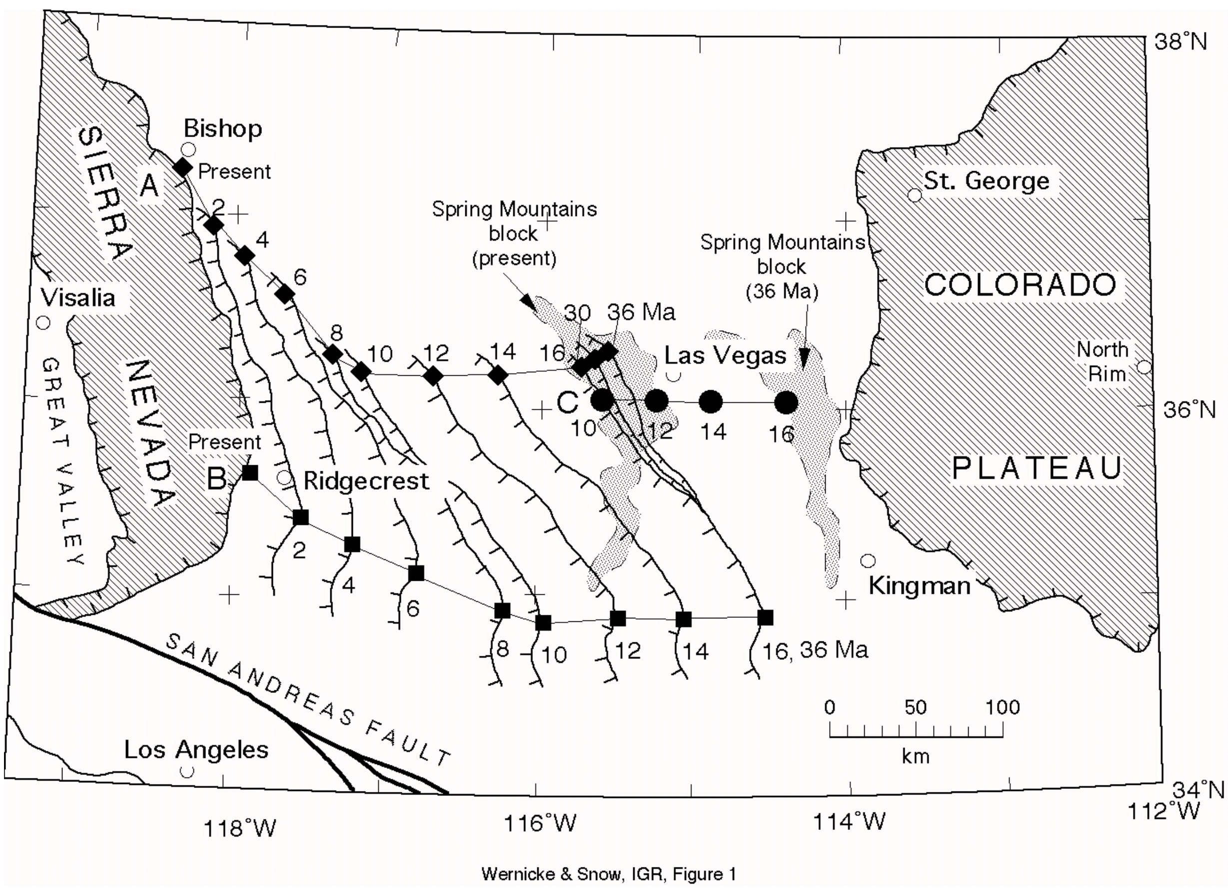


Figure 4: Map showing track of the Sierra Nevada block relative to the Colorado Plateau, based on geological reconstruction of Mesozoic and mid-Tertiary geological features. Diamonds and squares show positions of points A and B, respectively, on the Sierra Nevada block at the times shown, in millions of years. Circles show the approximate position of point C in the Spring Mountains relative to the Colorado Plateau (from Wernicke and Snow, 1998).

2.2 Earthquake Physics

Earthquakes are by far the most conspicuous manifestation of plate boundary deformation. The powerful shaking and destruction that accompanies large earthquakes is a constant reminder of the tremendous elastic energy that is stored in plate boundary deformation. The spatial distribution of seismicity is reasonably well understood. It clearly marks those areas where plate strain is concentrated, such as along the San Andreas Fault system **(Fig. 2a)**. But what controls the timing of earthquake occurrence? As in the previous section, we benefit from the knowledge that the relative plate motion ultimately responsible for earthquakes is essentially constant in time; but what we observe is a complex space-time pattern of seismic activity. What is the relationship between the two? Plate motion supplies some temporal information in that the total moment release rate (proportional to the product of slip-rate and fault area) of earthquakes should, on average, scale with the relative plate velocity, if all slip is seismic. This reasoning has led to the concept of an earthquake cycle, where elastic strain is gradually accumulated at a rate controlled by the plate motion, released in an earthquake, and then reaccumulated. While this idea is a useful starting point for understanding earthquake occurrence, the resulting simple picture of quasi-periodic earthquake occurrence on isolated faults does not adequately account for the much more complex sequencing of earthquakes, and is thus an incomplete physical description. It does not, for example, include the interactions between faults or earthquakes, or more generally the continual redistribution of stress and strain within the plate boundary system, nor does it contain a description of the specific physical process by which earthquakes nucleate.

In some cases, these additional factors are dramatically illustrated by patterns of earthquake activity. For example, the recent devastating M=7.4 Izmit earthquake in Turkey marks the latest in a 60-year sequence of major earthquakes that define a westward migration of seismic activity along a 500-km-long segment of the strike-slip North Anatolian Fault. It is likely that the stress changes due to previous earthquakes are important in causing the succeeding quakes, but what controls the time between mainshocks? The elastic stresses are transmitted nearly instantaneously and have been invoked to account for correlated earthquake activity (e.g., Reasenberg and Simpson, 1992; Stein et al., 1992). Yet, the time between successive earthquakes ranges from years to decades. A number of processes could contribute to the time lag: viscoelastic relaxation of the lower crust, postseismic fault slip, pore-fluid flow, or rate-dependent friction on the triggered fault. We have also seen dramatic triggering of earthquakes by propagating seismic wavefronts from other earthquakes. The Landers earthquake of 1992 triggered seismicity throughout the western United States (Hill et al., 1993); the Izmit earthquake triggered seismicity in Greece (Brodsky et al., 2000). How does this triggering occur? Does the wavefront produce semi-permanent changes in the distant stress field? Foreshocks, small events that often precede major earthquakes, are indicative of a nucleation phase. Indeed, many of the recent large earthquakes in southern California, such as the recent 1999 Hector Mine earthquake, possessed a well-defined foreshock sequence. Is there an aseismic component to foreshock occurrence? Are foreshocks part of a more complex nucleation phase?

In each of these examples, the particular phenomenon has been inferred solely from patterns of seismicity. Yet, seismic data alone are insufficient to determine uniquely the underlying physical process. The accompanying aseismic strain, an integral part of the earthquake process, remains largely unobserved. PBO would provide data crucial to answering these questions, and improving our ability to forecast future seismic hazards. These aseismic processes that operate during the earthquake cycle take place on timescales from minutes up to hundreds or even thousands of years. While seismology provides critical data at the short times relevant to elastic wave propagation, and earthquake geology yields constraints on long-term time series and rates of elastic strain release, it is high-precision geodesy that can provide a detailed image of deformation during the intermediate stages of the earthquake cycle. These details are essential for understanding earthquake physics and the dynamics of active fault systems. Relevant processes that we expect to observe include the nearly steady plate-tectonic motions that drive earthquake activity; the strain accumulation and possible precursory deformation that leads to episodes of fault slip; the strain redistribution associated with the clustering of earthquakes in space and time; the post-seismic deformation that follows a major event; and the stress interactions within systems of active faults.

Observing earthquake-related transient deformation

A key observational requirement in the study of earthquakes is to observe transient deformation. While a rich phenomenology of aseismic behaviors is expected, and in some cases known to accompany earthquake occurrence, none of these transient processes has been observed well enough to be understood in terms of the basic physics of the Earth's crust. Theories of frictional fault slip suggest the existence of a nucleation phase preceding earthquake rupture, but existing monitoring networks are not sufficiently dense to test these theories definitively.

Existing but limited data from the San Andreas Fault system and elsewhere do, however, reveal very intriguing examples of transient phenomena, and thus suggest that a broad class of behaviors would be observed by PBO (**Fig. 5**). For example, post-seismic deformation following the Landers earthquake has been observed with three distinct time constants by three instrumentation types: 6 days by strainmeters (Wyatt et al., 1994), 50 days by GPS (Shen et al., 1994) (**Fig. 6a**) and 3 years by SAR (Peltzer et al., 1996; Massonnet et al., 1996). These various post-seismic transients are suggestive of multiple deformation mechanisms and show exciting evidence for both stress relaxation in the very shallow crust and ductile flow in the lower crust and upper mantle (Deng et al., 1998). This type of observation is not only vital for understanding earthquakes, but is also important for estimating the vertical distribution of rheology in the lithosphere. Two other examples illustrate the potentially broad range of transient behavior. The first is a slow earthquake (duration ~10 days) detected along the San Andreas Fault near San Juan Bautista, seen on two strainmeters, that was also accompanied by increased seismic activity (Linde et al., 1996, **Fig. 6b**). This is one of the best-documented slow earthquakes yet observed. The second is multiyear aseismic transient increase in the San Andreas fault slip-rate near Parkfield which was observed on an electronic distance measurement (EDM) network (comparable in precision to GPS), strainmeters, creepmeters, and was accompanied by an increase in microearthquake activity (**Fig. 6c**, see Gao et al., 2000, and references therein). This was not only a particularly well-documented transient, with observed synchronous variations in surface and subsurface strain, but also showed evidence for aseismic stress transfer along the fault. Results from the recently installed 1000-station continuous GPS network in Japan show examples of slow earthquakes (Hirose et al., 1999), in addition to post-seismic fault displacements with magnitudes comparable to coseismic movements (Heki et al., 1997). In addition, the 1994 Northridge earthquake shows afterslip on the mainshock rupture plane as well as triggered slip on neighboring faults. The distribution of aftershocks for this event, combined with the existing geodetic data, suggests that viscoelastic relaxation of the lower crust may control the timing and spatial distribution of aftershocks (Deng et al., 1999). However, network density is sparse, time resolution is very coarse, and interpretations are uncertain and non-unique.

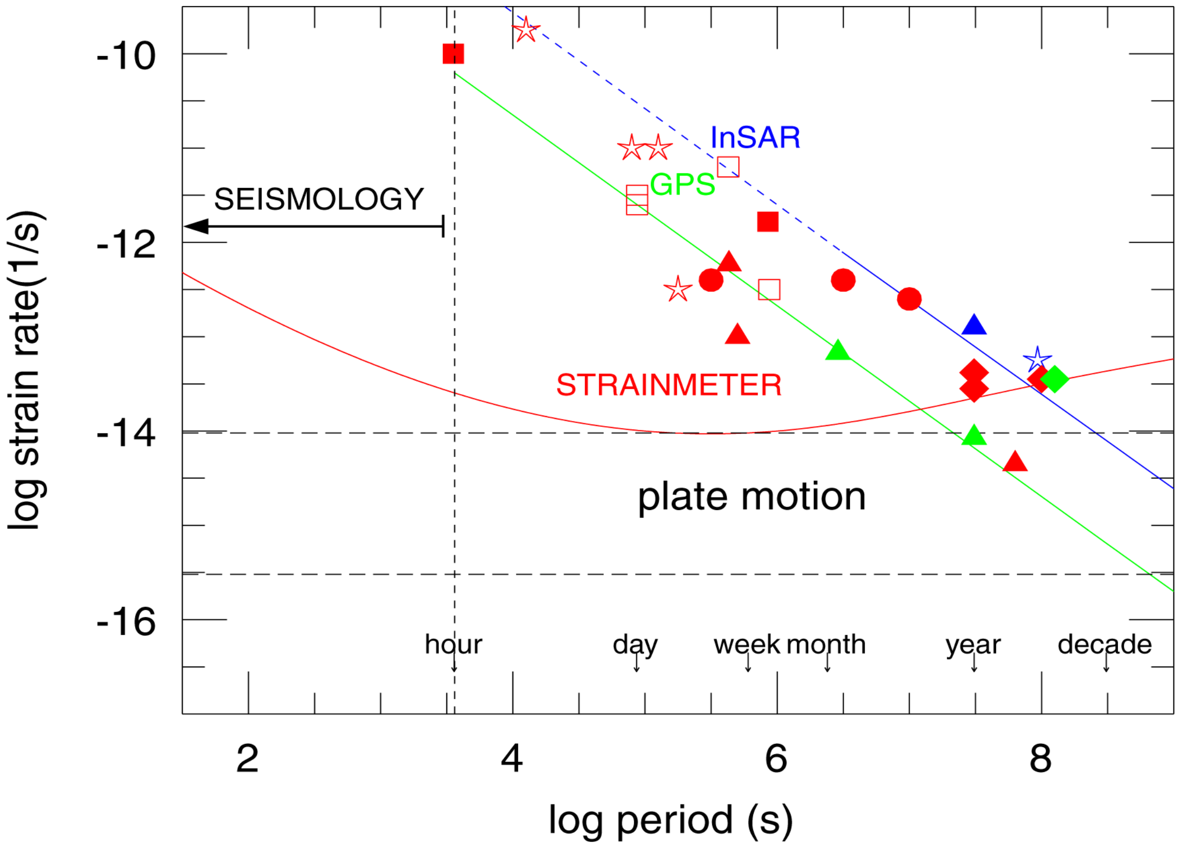


Figure 5: Proposed components of the Plate Boundary Observatory, and observed transients. Thresholds of strain-rate sensitivity (schematic) are shown for strainmeters, GPS, and InSAR as functions of period. The diagonal lines give GPS (green) and InSAR (blue) detection thresholds for 10-km baselines, assuming 2-mm and 2-cm displacement resolution for GPS and InSAR, respectively (horizontal only). GPS and InSAR strain-rate sensitivity is better at increasing periods, allowing, for example, the detection of plate motion (dashed line) and long-period transients (periods greater than a month). Strainmeter detection threshold (red) reaches a minimum at a period of a week and then increases at longer period due to an increase in the amplitude of the Earth noise spectrum. At long periods (months to a decade) GPS has greater sensitivity than strainmeters by one to two orders of magnitude. At intermediate periods (one week to months) sensitivities are comparable, and at shorter periods (seconds to a month) strain sensitivity is one to three orders of magnitude greater than GPS. Combined use of both instrument types provides enhanced sensitivity for detection of transients from earthquakes to volcanic eruptions to plate motion. Also shown are several types of transients observed by strainmeters (red), GPS and equivalent (green), and InSAR (blue): Post-seismic deformation (triangles), slow earthquakes (squares), long-period aseismic deformation (diamonds), preseismic transients (circles), and volcanic strain transients (stars) (from Silver et al., 1999).

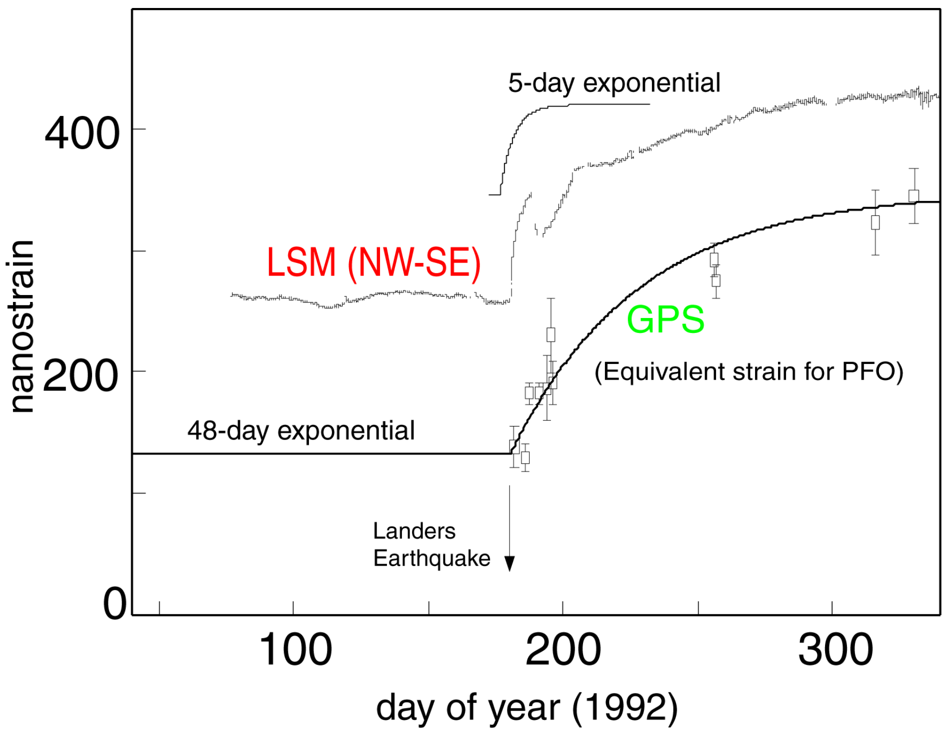


Figure 6a: Post-seismic deformation for the 1992 Landers earthquake from strainmeters (LSM) and GPS, illustrating roughly 6-day and 50-day time constants, respectively (after Wyatt et al., 1994). This illustrates that post-seismic deformation may have several characteristic time constants.

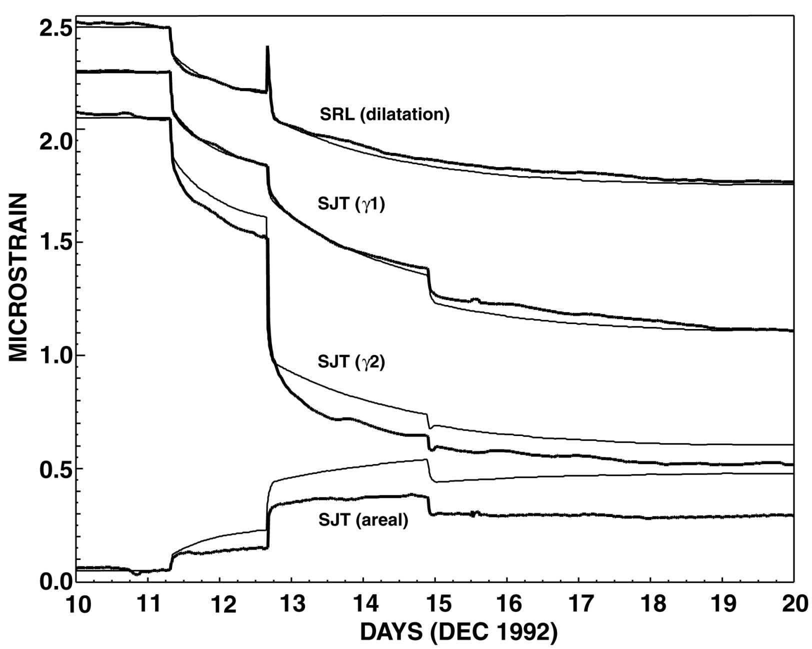


Figure 6b: Slow earthquake (10-day) detected along the San Andreas Fault at San Juan Bautista (south of Bay Area) by two strainmeters, and accompanied by elevated seismicity (Linde et al., 1996). This is one of the best-documented slow earthquakes yet observed

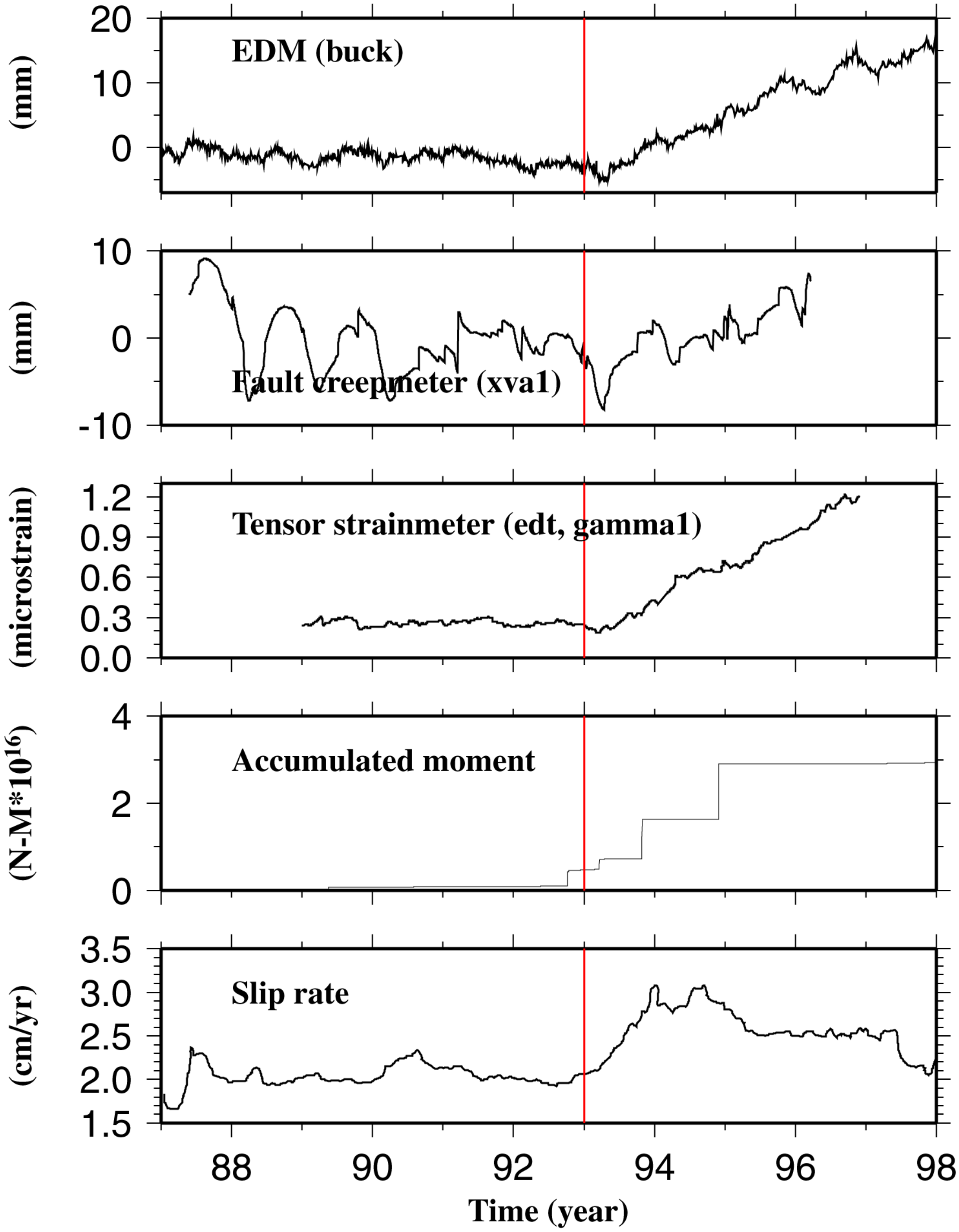


Figure 6c: Multiyear aseismic transient in San Andreas Fault slip at Parkfield, observed by electronic distance measurement (EDM) (GPS equivalent), strainmeters (Gwyther et al., 1996), creepmeters, an increase in estimated slip rate from repeating microearthquakes (see Nadeau and McEvilly, 1999), and accompanied by a general increase in seismic activity (accumulated moment). All indicators suggest an acceleration of slip on the San Andreas Fault at Parkfield, and aseismic stress transfer (Gao et al., 2000).

The general problem that we seek to address through the deployment of PBO is to more fully understand the physics of earthquakes by observing the aseismic deformation field. We list below three basic questions along with the needed observations.

Basic questions

• *What controls the space-time pattern of seismicity?*

Is seismic activity correlated or essentially random? We have already noted some cases of proposed correlated seismic activity, such as the sequence of Turkish earthquakes. There are many others. Rydelek and Sacks (1988) noted a correlation between landward and trenchward events in Japan that they attributed to post-seismic viscoelastic stress relaxation. Pollitz et al. (1998) have suggested that post-seismic relaxation from several major earthquakes in Alaska in the 1950's and 1960's and subsequent southward migration of stress and strain produced a corresponding southward migration of seismic activity along the San Andreas Fault system. Press and Allen (1995) proposed that changes in tectonic activity in the Basin and Range are causally related to the pattern of seismicity and style of faulting along the San Andreas Fault system. Finally, there are many cases of apparent clustering of activity, such as in southern California over the last decade, with the occurrence of the Landers, Joshua Tree, Big Bear, Northridge, and most recently, Hector Mine earthquakes.

Fault-fault interactions may represent a potentially important form of correlated seismic activity. A good example is the interaction between the San Andreas and Hayward Faults in northern California. Creep rates along the southern end of the Hayward Fault slowed significantly following the M=7.1 1989 Loma Prieta earthquake (Galehouse, 1995; Lienkaemper et al., 1997; Bürgmann et al., 1998). Such behavior indicates that fault-fault interactions occur within the Bay area, presumably through the static stress field generated by the Loma Prieta event, which also affected the microseismicity pattern in the region (Simpson and Reasenberg, 1994; Gross and Bürgmann, 1998). We know that this stress field has been relaxing with time (Savage et al., 1994; Segall and Bürgmann, 1997) and may further alter motions on the Hayward fault and regional seismicity. Only continuous observations of crustal deformation can adequately address how fault behavior changes with time and can elucidate the timescales of processes that occur throughout the earthquake cycle.

The region below the seismogenic zone may play a critical role in fault-fault interactions. Is the seismogenic zone underlain by a homogeneous ductile layer or does the fault extend aseismically below active fault zones? Surface geodetic velocity measurements provide some constraints that may help to distinguish between these two possibilities. In addition, seismic imaging can play a central role by mapping faults at depth. For example, a vertical offset in the Moho directly beneath a fault would suggest the existence of a crustal fault with a significant aseismic extension (e.g., Henstock et al., 1997).

Directly observing the interseismic strain accumulation field is central to understanding the space-time distribution of seismicity. This deformation field is a proxy for the elastic strain energy change that drives crustal earthquakes, and is thus a first-order predictor of future seismic potential. The space-time distributions of the seismic and aseismic fields are intimately linked. While some information exists about the interseismic deformation field along the San Andreas Fault system, the data are neither sufficiently precise nor well distributed to directly observe this redistribution of strain.

Studying the space-time variation of seismicity benefits greatly from the long-term temporal framework provided by geologically-based constraints on seismicity. This extension of the seismicity timescale is possible through paleoseismological investigations of active fault systems. Using this technique on a number of trenches, it was possible, for example, to document the last ten events on the Pallett Creek segment of the San Andreas Fault. The age brackets of these events suggest approximately one surface-rupturing earthquake every 132 years, and tendency for temporal clusters of two to three events, separated by dormancy periods of 200-300 years (Sieh et al., 1989). Fault zones within the interior of the plate boundary zone are mostly short (ca. 50 km), consist of normal and strike-slip fault segments, and have slip rates and event frequencies an order of magnitude lower than major plate boundary transforms and subduction zones. Despite this difference, they account for some 20-30% of the total plate boundary strain budget and produce devastating earthquakes although at reduced frequency. Yet the relation of the intraplate system to the major plate boundary structures and volcanism, and the basic pattern of strain release within the system itself, while of great interest (Press and Allen, 1995), is largely unknown. Obtaining precise earthquake occurrence information on these structures is needed in order to observe the spatial and temporal pattern along and across the system.

• *How do earthquakes nucleate?*

A major objective of earthquake science is building a complete theory of earthquake nucleation. While there are several models and ideas (e.g., Dieterich, 1980; Brune, 1979) that predict the existence of an aseismic nucleation phase, strong constraining observations are absent. Expected nucleation signals are likely to be small in amplitude and short in duration, and have so far eluded detection. The major requirement for studying the problem is that we have instruments close to the impending earthquake, with high sensitivity in the short-period (seconds to a month) end of the geodetic band. The direct monitoring of subsurface strain within the seismogenic zone would be very valuable as well. Presently this can be done by using microearthquake activity as a proxy for aseismic deformation. Since we need the ability to detect small transient strain phenomena near a large earthquake, and because we cannot presently predict earthquakes, it is essential to have excellent spatial and temporal resolution over historically seismogenic areas. This means that a central focus for geodetic deployments must be the San Andreas Fault system, especially its most seismogenic segments. High-resolution microearthquake locations are also needed in these same regions. Observations of correlated subsurface and surface transients would be especially valuable in the study of the nucleation phase.

• *How can we reduce the societal risk of earthquake hazards?*

One of the significant societal benefits of a greater understanding of earthquake physics is the expected reduction in earthquake risk. Our primary approach to reducing earthquake risk through mitigation relies on long-term forecasts of where and when earthquakes are likely to occur. A critical task here, both from a fundamental and practical viewpoint, is to determine the fraction of fault slip that is seismic, and to identify those parts of the plate boundary that are expected to slip seismically. Geodesy provides critical information here that can be used in conjunction with seismic catalogs and paleoseismology. It can provide an independent estimate of the magnitude of strain accumulation, and spatial variations in deformation can be used to identify locked, and presumably seismogenic, sections of the fault. Various segments of the plate boundary zone apparently exhibit different behavior. Along the San Andreas Fault system, most of the plate motion appears to be released in large earthquakes. In Alaska, the seismic fraction is probably closer to a half, and it has been possible to identify a locked section of the plate interface, using existing GPS data (Freymueller et al., 1999). The Cascadia subduction zone is particularly intriguing, in that there are no large recorded events, initially suggesting that the 4 cm/yr of plate convergence is entirely aseismic. The last decade, however, has seen paleoseismic evidence for very large, infrequent events (500-year recurrence interval) suggesting a large fraction of the slip may indeed be seismic. The GPS data for Cascadia, revealing a strongly deformed overlying plate, are also consistent with a locked interface, strengthening the case for a large seismic hazard in this region. Finally, geodesy is particularly valuable where the earthquake history is not well known. For example, initial GPS data across the Wasatch Fault zone (and near the Yucca Mountain, Nevada, nuclear waste repository) imply strain accumulation that is much faster than implied by earthquake history and paleoseismology. If so, the seismic hazard may be considerably higher than previously assumed. (Martinez et al., 1998).

Ideally, such long-term forecasts would be supplemented by short-term predictions that have been so valuable for other natural hazards. Can we observe a nucleation phase? Are there preseismic transients, in addition to foreshocks, that could form the basis for predictive capability? The hypothesis that aseismic deformation might precede significant earthquakes, and be detectable by geodetic instrumentation, requires careful study. While there have been tantalizing preseismic signals to earthquakes over the years, few have been sufficiently well-documented to be definitively confirmed or rejected. This is largely due to the absence of instrumentation in the zone of the impending earthquake. Because earthquakes cannot presently be predicted, we require sufficient instrumentation in historically seismogenic areas so as to 'capture' large events and observe the preseismic strain field with the greatest possible sensitivity. With the deployment of PBO, we should thus be able to thoroughly test this earthquake predictability hypothesis. Whatever its outcome, this test constitutes a critical step in our continuing quest to reduce society's vulnerability to devastating earthquakes.

Required Observations

- *Geodetic measurement of surface deformation in historically seismogenic areas to understand the character and mechanisms of transient deformation related to earthquake occurrence. This requires high-spatial-resolution determination of the steady state strain accumulation, as well as the ability to measure strain transients with time constants ranging from minutes to decades.*

- *Complementary observations of deformation at seismogenic depths by the continual observation of microearthquake activity and other strain indicators.*

- *An extended history of earthquake occurrence over many earthquake cycles, obtained from geological measurement, to understand more fully the causes of earthquake occurrence.*

- *Systematic imaging of faults and other earthquake-related structures through the use of seismic tomography.*

2.3 Magmatic Processes

Volcanism occurring in the Pacific-North American plate boundary zone is a consequence of several processes: subduction, as in the Cascadia and Aleutian arcs, crustal extension, as in the Basin and Range extensional province, and hotspot activity, as in the Yellowstone area. In each case, magmatic processes are intimately connected to the overall tectonic driving mechanism, so that understanding these links is critical to understanding the underlying processes and driving forces. The plate boundary deformation zone under study offers a natural laboratory for studying the magmatic processes associated with subduction: Alaska possesses 42 active volcanoes, and the Cascades have 17. The Snake River Plain-Yellowstone region is an ideal location to study the complex interplay between tectonic, seismic, and magmatic processes.

Like earthquakes, there are a wide range of volcanic transients. For example, the Yellowstone caldera shows evidence for decade-long deflation of the caldera as well as deformation related to slip on the nearby Hebgen Lake Fault (**Fig. 7a** from Smith et al., 1999). The ascent and migration of magma and associated fluids and gases are necessary preludes to volcanic eruption. There are clearly pre-eruption strain transients that have preceded eruptions. For example, at Hekla, in Iceland nearly identical pre-eruptive signals were obtained from strainmeters (**Fig. 7b**), both for the January 1991 eruption (Linde et al., 1993) and for the recent February 26, 2000 eruption (A. T. Linde, personal communication). At Kilauea (Owen et al., 2000, **Fig. 7c**), a pre-eruptive transient was measured with continuous GPS. There have also been several other kinds of transients more generally related to the movement of magma recorded at Long Valley, California (Linde et al., 1994; Langbein et al., 1995), Arenal volcano, Costa Rica (Hagerty et al., 1997), Montserrat (Mattioli et al., 1998) and during an eruption of Izu-Oshima, Japan (Linde et al., 1996). Magma transport depends upon tectonic conditions and itself influences faulting and earthquake occurrence. However, the rate of magma ascent, its residence time and location in the crust, and the processes that lead to eruption are either poorly understood or totally unknown. Geodetic monitoring of the surface deformational patterns and their temporal evolution permit magmatic sources of deformation to be located and tracked with time.

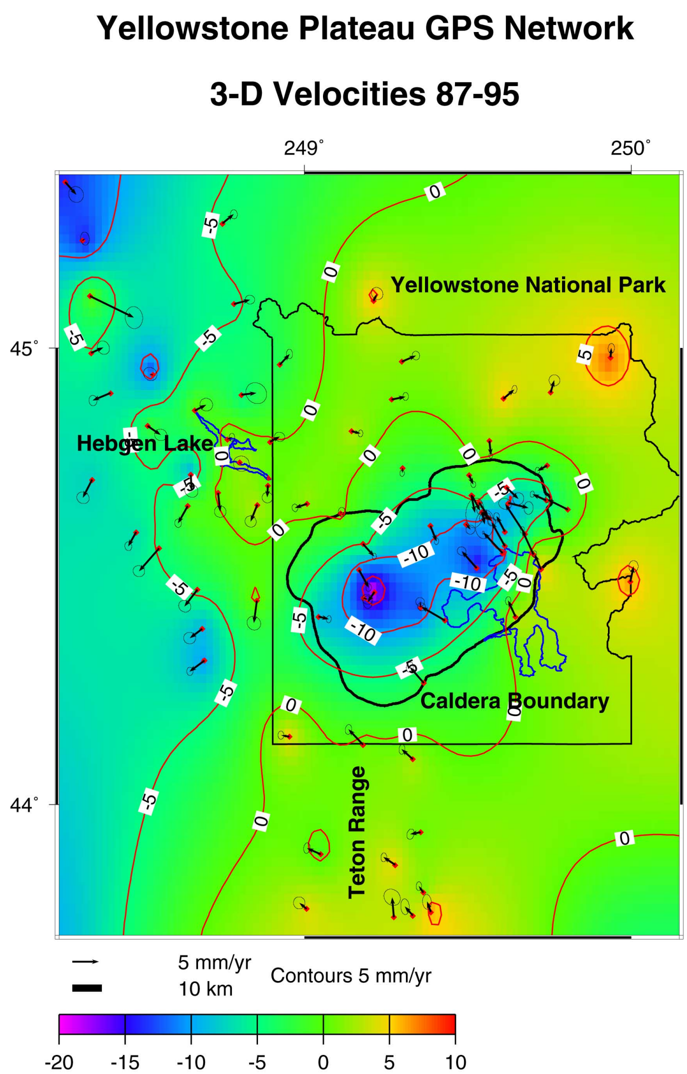


Figure 7a: Dramatic deflation of Yellowstone Caldera from 1987-1995 from a GPS survey, performed by Smith et al. (1999).

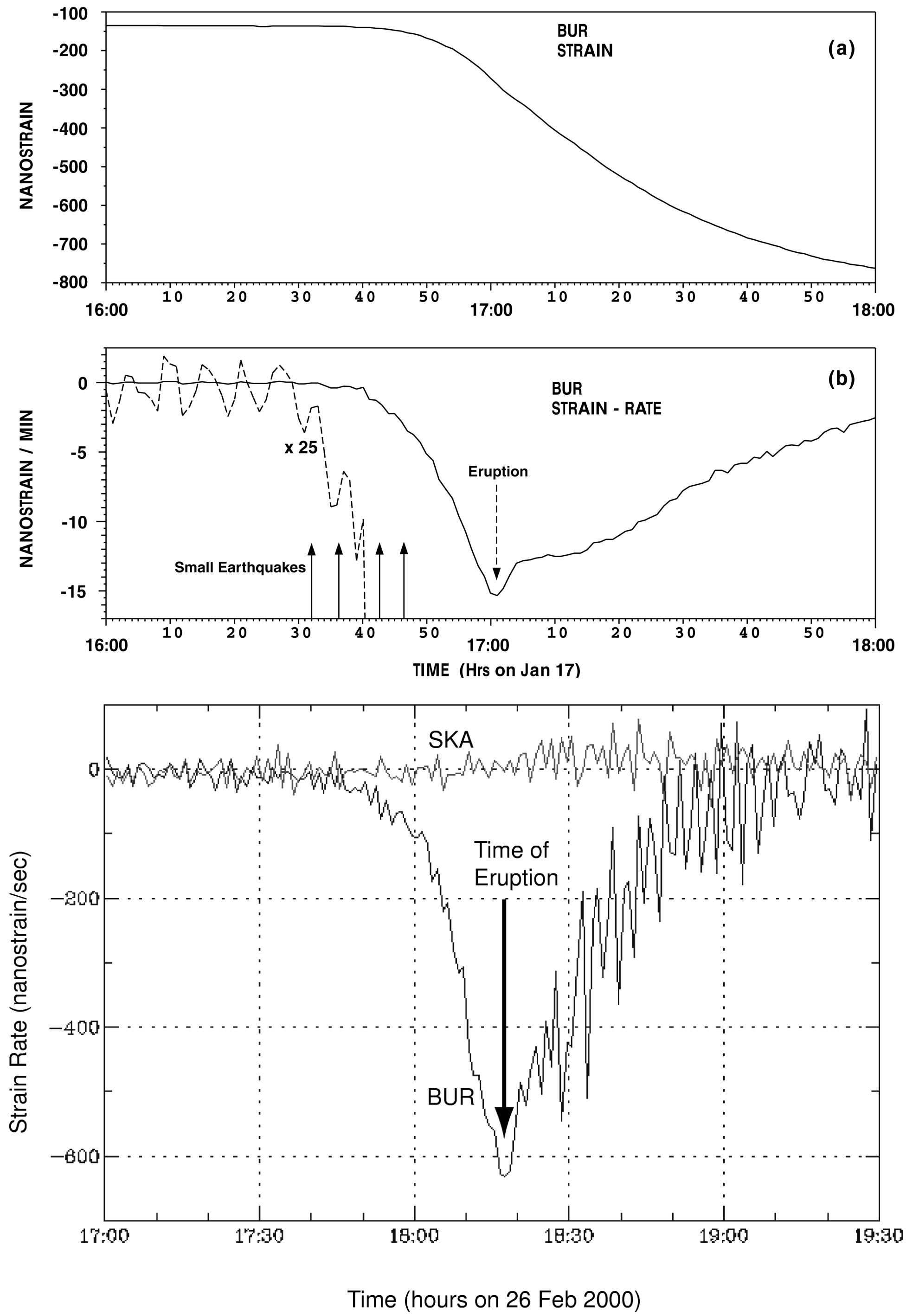


Figure 7b: Pre-eruptive strain changes before two Hekla (Iceland) eruptions (from Linde et al., 1993 and A. T. Linde, personal communication). Borehole strain data for the eruption of January 1991 (top and middle). Two hours of strain data (top) and strain rate data (middle) from BUR, a borehole strainmeter site 15 km from Hekla. The dashed strain rate curve is magnified 25 times. Minimum in strain-rate gives the inferred time of surface breakout (dashed arrow), known from other observations to be between 17:00 and 17:05. Also shown by solid arrows are the times of the first located earthquakes associated with the eruption. The eruption begins with the initiation of dike formation about 30 minutes before the eruption and the small earthquakes (M~1) occur after the magma starts to move. Bottom, a nearly identical record of strain rate from BUR for the recent Hekla eruption of February 26, 2000 also showing a pre-eruptive strain change.

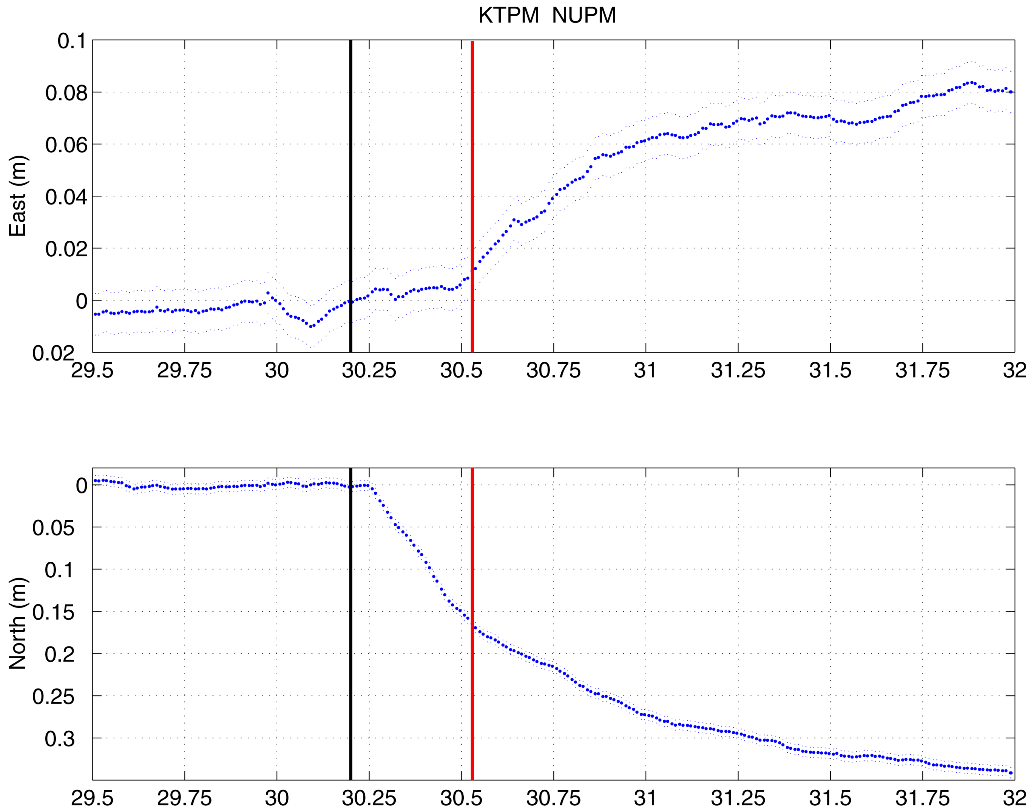


Figure 7c: Pre-eruptive velocity changes from GPS, studied by Owens et al. (2000). Extension across the East rift zone of Kilauea volcano (Hawaii) as a function of time from continuous GPS measurements. Distance between continuous stations NUPM-KTPM is shown for the period of eruptive activity at Napau Crater. Timescale is Day of Year, 1997. First vertical line indicates the onset of tremor at Napau Crater and the second the onset of the eruption. Southward motion of KTPM shows extension of the rift.

Sufficiently precise monitoring networks can be used to define volcanic source geometry, follow it in time, and clarify the processes responsible for magma movement and volcanic eruptions. In most of the plate boundary zone, however, current coverage is sparse or non-existent. Minimally adequate continuous monitoring is being initiated at Kilauea volcano on the island of Hawaii, and at Long Valley, and is essentially absent elsewhere. InSAR 'snapshots' of volcano deformation are beginning to define geometry and locations of sources of deformation at a fraction of the world's active magmatic systems, but coverage and repeat frequency are severely limited by the currently available satellite radars. However, the emerging evidence from InSAR mapping of many volcanoes shows a surprising variety of deformation sources, making them rich targets for intensive geodetic monitoring. Results from such networks will enable mapping of the deformation sources and will clarify the relations between these sources and the processes of magmatic migration, differentiation, and degassing. Such understanding will be essential for informed assessments of volcano hazard and eruption prediction.

Basic questions

- *What are the dynamics of magma rise, intrusion, and eruption?*

Magma forms whenever temperatures in the mantle exceed some threshold value, specifically the solidus temperature for peridotite. For example, the solidus temperature may be lowered by addition of water at subduction zones. The partial melting product of peridotite, basalt, is less dense than the surrounding mantle, and thus rises due to buoyancy forces until it reaches a level of neutral bouyancy, typically in the middle or upper crust, or continues upward and erupts as basalt or basaltic andesite. Once magma reaches the middle or upper crust, its motion can be tracked fairly well with surface deformation data (Dvorak and Dzurisin, 1997). Magmas that "pond" for a significant period in the crust undergo fractional crystallization and assimilation of lower melting point crustal material, usually more silica-rich in composition, thus forming more fractionated andesitic to rhyolitic lavas, or intrusive bodies with similar chemistry. The factors that control whether a magma erupts at the surface or intrudes and cools within the crust include initial composition and temperature, rate of formation, rate of rise (itself a function of temperature, composition and rheology), and the nature of surrounding crust. Variables that control composition include starting composition of the mantle, percentage of partial melting, and the nature and amount of assimilated crustal products. Compositional factors that are especially critical to eruptive style include silica and volatile (CO_2, SO_2 and H_2O) content, since these control the viscosity and the pressure history of the rising magma (volatiles such as CO_2 exsolve strongly as pressure is reduced, and thus are a critical factor for explosive eruption). To the extent that surface deformation patterns in space and (especially) time will be influenced by the viscosity of moving magma, detailed deformation data could in principle be used to constrain the viscosity and hence the composition of magma before it erupts. Limited available data suggests that time constants for deformation episodes in basaltic shield volcanoes such as Hawaii and Iceland are typically a few hours to a few days (consistent with relatively rapid motion of low viscosity basaltic magma), while deformation time constants for silicic calderas such as Long Valley may be several months.

- *What is the size and shape of magma reservoirs and conduits and how do they constrain the dynamics of magma flow and deformation?*

Source geometry (size and shape) are hard to constrain, but they are important because: (1) the chamber geometry influences pressure history as the magma chamber drains, (2) other things being equal, large chambers have the potential to erupt in larger explosive eruptions, (3) errors in assumed model geometry lead to errors in inferred depth, and (4) the geometry has a significant effect on stress changes which are in turn related to seismicity associated with magma movement. To address this problem, we require detailed surface deformation measurements, the monitoring of microearthquake activity, (as a measure of subsurface deformation), and seismic imaging.

- *How do the temporal and spatial scales of deformation vary with eruptive style and magma composition?*

The basic information for answering this question is the direct observation of transient deformation associated with magma ascent. At this time we have far more data for basaltic shield volcanoes than we do for more explosive andesitic stratovolcanoes or silicic calderas. It is critical to obtain deformation data at different types of volcanoes if we are to understand the links between surface deformation, magma type, and eruptive style and potential.

• *Can we characterize deformation that leads to an eruption, and predict eruptions with high confidence?*

Detectable strains precede most eruptions for which monitoring data exist. However, there are examples of volcanic deformation that have not lead to an eruption. There are also examples of small eruptions which appear to have no significant strain, although this might simply reflect the location of instruments too far from the center of activity, or instrument sensitivity that is too low. Our current understanding of the physics of magma rise and eruption suggests that we ought to be able to predict all major eruptions and most minor ones, given sufficient data. The dense instrument array we propose to deploy on selected volcanoes within the Pacific North American plate boundary zone will allow this question to be answered definitively, and will improve our understanding of magma dynamics. Most volcanologists believe that for well-monitored volcanoes, eruptions can be predicted with very high confidence, with at least days, and perhaps months, of warning. The basis for this confidence is outlined below.

Volcanic hazard assessment begins with identification of volcanoes that have been active recently (on a geological) and thus have the potential for future activity. Long term forecasts can then be constructed based on the geologic record and historical activity. A good recent example is Cerro Negro volcano in Nicaragua, where Hill et al. (1998) successfully predicted a future eruption based on historical activity for the time window 2000 AD±1 year (the eruption actually occurred in August 1999).

Short-term predictions are made based primarily on ground deformation, seismicity, and gas emissions. Gravity information can also be useful for assessing the density of new material entering the system. All volcanologists agree that a variety of data are required to generate high confidence eruption forecast/warnings, because of the broad range of volcanic behavior in both space and time. For example, as noted above, borehole strainmeters have demonstrated sensitivity sufficient to “image” magma motion on its way to the surface, and have detected signals 30 minutes in advance of a basaltic eruption at Hekla volcano in Iceland (Linde et al., 1993). But the magma that erupted at Hekla must have accumulated at deeper levels in the crust over a longer period (months or years) where the noise levels for strainmeters are higher. Most volcanologists believe that such slow accumulations of magma can also be detected, and thus give very useful longer term warning, e.g., by GPS. At Long Valley, ground deformation was observed several months before the onset of seismicity that likely heralded magmatic motion at depth, although an eruption did not occur (Langbein et al., 1993). It is possible that had GPS or other long-period deformation monitoring instruments been deployed at Hekla, slow accumulation of magma in a crustal magma chamber might have been detected weeks or months before the eruption. GPS, leveling, and tilt data showed clear eruption precursors roughly one week prior to a recent eruption in Japan (Aoki et al., 1999).

GPS is an excellent tool for measuring ground deformation. It has several advantages over conventional EDM and leveling, and is becoming a powerful tool for volcano monitoring. In the semi-permanent station mode, GPS gives high precision, automated, continuous deformation data in all weather, in a cost-effective manner and with little risk to personnel (Dixon et al., 1997). When combined with borehole strainmeters (which have much higher short-period sensitivity but less long-term stability), GPS can provide critical data for eruption prediction. This is the ground deformation instrument suite we propose for PBO. Together with seismic and gas emission data, this forms a comprehensive volcano monitoring system, capable of medium to long term eruption forecasts, and accurate short-term predictions. More importantly, this system will lead to greatly improved understanding of magmatic systems.

Required observations

- *The observation of transient deformation, both surface (geodetically) and subsurface (through seismicity), is needed to track the movement of magma and associated fluids from depth to the surface. This requires the measurement of deformation with high spatial resolution and the ability to measure strain transients with time constants ranging from seconds to decades.*

2.4 General Observational Considerations

In designing the PBO's capabilities, the basic observable is the strain rate or the velocity-gradient field. Thus, strain rate sensitivity both as a function of wavelength and period are the first-order considerations. One important observational goal is to recover the **decadal strain rate field** with sufficient spatial resolution to observe how it is being partitioned within the plate boundary zone. The existing geodetic constraints on the decadal field provide a good guide to the observational requirements. Existing deformation data (GPS and VLBI) for part of western North America (**Fig. 2a**) indicate the relative size of some of the steady state signals associated with different features of the approximately 50 mm/yr steady state PBZ motion (e.g., Bennett et al., 1999). These data show that velocity accuracy of about 1mm/yr accuracy will be needed. The primary feature, the overall strike-slip deformation along the San Andreas Fault System, takes up about 35 mm/yr of this motion (Shen et al., 1996; 1997). Resolving the motion to perhaps the 1 mm/yr level is adequate to show how the strike-slip motion is partitioned between the various known faults (Bourne et al., 1998). In addition, understanding finer structure, such as the effects of the locked thrust faults in the Los Angeles basin responsible for the Northridge earthquake, requires similar resolution, because the fault location and geometry are not well known, and their surface deformation is small. A similar conclusion emerges from consideration of Basin and Range tectonics. The general pattern is about 12 mm/yr of approximately NW-SE extension, most of it focussed in 40-100 km wide zones (Thatcher et al., 1999; Dixon et al., 2000). Resolving the details of this extension will require data with an accuracy of 1 mm/yr or better and station spacing as small as 20 km. We expect the situation to be similar throughout much of the plate boundary zone. For example, there is an overall ~10 mm/yr deformation of the overriding North American plate by the subducting Juan de Fuca plate (Dragert et al., 1999; Goldfinger et al., 1999), but this pattern contains important spatial details that need to be known to higher precision to constrain the details of the subduction process. Thus, resolving the decadal field with sufficient spatial resolution and sensitivity will require 1 mm/yr velocity accuracy over distances as small as ~20 km.

The second important observational goal, and the one for which we have much less observational experience is recovering the **time-dependent component** of the strain rate field. As noted above, deformation transients are central to furthering our understanding of earthquakes, magmatic processes, and the rheology of the lithosphere and asthenosphere. Precursory transients to major volcanic eruptions and earthquakes, if they can be identified, would be of great societal value to improve our preparedness for these natural hazards. To study deformation associated with these short-term processes, high temporal resolution is thus crucial. As with the decadal field, an examination of published transients **(Fig. 5**) provides a guide to what is needed. Available observations, such as those noted above, demonstrate the existence of such transients over the entire minute-to-decade range for both earthquakes and magmatic events. Thus a reasonable design goal is in areas of high strain rate related to either seismic or magmatic activity, to have uniform strain rate sensitivity over this range that corresponds at least to the average yearly strain rate in the region.

3.0 PBO Deployment Strategy

Achieving the scientific goals enumerated in Section 2 requires a substantial new deployment of geophysical instrumentation to focus with unprecedented precision on critical, but thus far poorly understood, earthquake, volcanic, active tectonic, and geodynamic processes. The enormous range of spatial and temporal scales spanned by these processes mandate field deployments of a variety of instruments at a range of geometries. The PBO will necessarily include both widely distributed geodetic networks (‘backbone arrays’) to map out broad patterns of deformation rate at the scale of the plate boundary, and dense local networks (‘cluster arrays’) to focus on specific earthquake and magmatic processes.

Deployments will be preceded by an initial design phase, during which site characterization and evaluation are made, and simulations are done to choose instrument locations and determine the appropriate mix of instrument type necessary to provide optimum imaging of the targeted process. Included within the design phase will be examination of cluster areas to determine which are best suited to address specific scientific goals. The San Andreas Fault system and several active magmatic systems (e.g., Yellowstone Caldera, Mt. Ranier) are certain to receive detailed attention via cluster deployments. Other areas, such as within the Cascadia subduction zone, Alaska, and the Basin and Range province, are likely to receive attention as well. Precise siting decisions, however, will await completion of the PBO design phase. In addition to site characterization and simulation studies, the design phase will include several workshops during 2000-2001 where candidate deployments will be proposed and discussed. Workshops have, for example, proven very useful for designing the southern California Integrated GPS Network (SCIGN). Final deployment decisions will be made on the basis of scientific merit and compatibility with PBO objectives.

3.1 Geodetic Instrumentation

Before focusing on deployment strategy, it is necessary to consider the kinds of instrumentation available to achieve our goals. A thorough discussion of candidate instrumentation is given in Appendix A, so only the salient features are summarized here. A basic conclusion is that the design goals of PBO can be met only with an integration of existing deformation-measuring technologies. Thus, the PBO will combine several techniques that differ in spatial and temporal sampling, strain sensitivity, and cost. A combined strategy is required both on observational grounds and because integrating diverse data types offers more powerful constraints on tectonic processes (e.g., Pollitz et al., 1999). For the PBO, a mix of GPS, strainmeters, and InSAR observations are needed to exploit their complementary strengths. As seen in **Fig. 5**, at the long-period end of the range of operative timescales (beyond one month period) GPS is particularly well suited to study decadal motion and its spatial variations, as well as large regional transients, such as the postseismic deformation from major earthquakes. Strainmeters, in contrast, can sample local spatial scales and have excellent sensitivity (2 to 3 orders of magnitude superior to GPS) in the short-period geodetic band of minutes to a month. Such instruments have dominated observations of strain transients to date. InSAR provides an excellent means of surveying deformation over broad areas. It is also capable of tens of meters spatial resolution at monthly or greater intervals. InSAR has proven to be a powerful tool to characterize the large-scale postseismic deformation from earthquakes, but it can also resolve such small-scale deformation features as shallow creep associated with earthquakes and postseismic and interseismic deformation. Strain-rate sensitivity of InSAR, however, is somewhat lower than GPS, and it is more susceptible to systematic errors caused by atmospheric variations. A comprehensive characterization of plate-boundary-related deformation demands a close integration of these three types of surface geodetic techniques.

3.1.1 GPS

Regarding the utilization of GPS, there are two possible approaches to making the observations. Measurements can be done in survey-mode (SGPS), where sites are occupied periodically for short intervals, or by a permanent network of continuously recording GPS receivers. Although the equipment costs are significantly higher, continuous GPS (CGPS) can provide significantly more precise data for a variety of reasons, including avoidance of errors associated with setting up equipment, denser temporal sampling, which better characterizes time dependent error sources, and the use of a more stable geodetic monument. Furthermore, CGPS networks are essential for monitoring the transient deformation signals that are central to PBO goals. In contrast, SGPS sites give less precise velocities and generally poor temporal resolution, but are individually less expensive. For example, a CGPS site costs up to $50K to install and operate for a year, but a SGPS site can be established and surveyed twice for about $1K and so such sites are useful for densifying coverage between CGPS ‘backbone’ stations and cluster deployments. This capability is valuable in a region as large as the western North American plate boundary zone, where even with one thousand CGPS stations many active regions would be sparsely sampled. The mix of GPS observations depends on tradeoffs of the cost versus precision and sampling tradeoff for the problem under investigation. A variety of strategies combining the two are under discussion (e.g., Thatcher, 1999). Nonetheless, the main focus of the PBO is on precise measurement of both steady state and transient deformation signals, so CGPS networks remain at the core of the deployment strategy. Survey-mode measurements are proposed to provide capabilities and flexibility not supplied by the fixed networks, including supplying needed density in unmonitored regions and quick deployment following earthquakes or detected strain transients.

3.1.2 Strainmeters

As described in Appendix A, strainmeters are very sensitive to deformation in the period range of less than a month, and are therefore the instrument of choice for a variety of short-term transients that may be most closely associated with earthquakes and magmatic events. Borehole strainmeter (BHS) sites are projected to cost $100K each ($75K if colocated with a CGPS site, as is anticipated), and include three components of (horizontal) strain, two components of tilt, and a three-component seismometer. In volcanic applications, shallower, less sensitive tiltmeters have proven very useful. In addition, borehole seismometers provide a means of monitoring microearthquake activity, which provides an important measure of subsurface strain. Because strain signals in the elastic Earth fall off with the inverse cube of the distance, it is essential for the strainmeters to be placed near potential areas of transient deformation to maximize strain rate sensitivity to the movement of magma or the nucleation processes of earthquakes. For optimal detection, they should be placed at distances less than or comparable to the depth of the transient signal. For optimal resolution of transient processes, it is also necessary to have instruments out to a distance of at least twice the depth of the transient process. One consideration in the case of strainmeters is that such instruments can respond not only to tectonic signals generated but also to local, non-tectonic processes near the instrument. Thus it is necessary to have clusters of instruments so that they can be analyzed for spatial coherence. In addition, non-tectonic coherent sources of noise (such as barometric pressure, tides, or precipitation) can be addressed by direct measurement of the phenomena (e.g., barometric pressure) or modeling (e.g., tides).

3.1.3 InSAR

Interferometric Synthetic Aperture Radar (InSAR) can provide spatially continuous maps of surface change over broad areas. Hence, InSAR provides denser spatial sampling than the few kilometer spacing possible with even the densest practical GPS arrays. Moreover, InSAR samples a broad area, and hence can provide initial epoch data prior to an earthquake or volcanic eruption without requiring the foresight (or luck) to target a specific locus of future tectonic activity, as long as data were recorded. InSAR's primary limitations are the cost of a dedicated satellite mission and the possible limitations imposed by image decorrelation.

3.1.4 Integrated PBO Instrumentation

In summary, the core of the PBO will consist of continuous GPS stations, with high-stability monuments, combined with strainmeters. We envision survey-mode GPS playing an important but secondary role for the densification of the continuous GPS network in selected areas, for sampling candidate target areas of interest, and in response to transient events such as large earthquakes or volcanic crises. This recommendation derives from the scientific problems we seek to address. In particular, we seek information about important features of the deformation field characterized at the one mm/yr level in velocity, which can be better obtained (and in a shorter observation time) using CGPS data. Hence we envision a PBO CGPS network similar conceptually to, and building on, the present SCIGN, BARD, BARGEN, EBRY, and PANGA networks.[1]

With these considerations in mind, we describe the deployment strategy appropriate for achieving each of the PBO science goals and observational requirements described in Section 2.0.

3.2 Deployment Needs by Observational Requirement

3.2.1 The Decadal Deformation Field

The primary observational requirement for understanding **plate boundary dynamics and evolution** is a detailed map of the present-day decadal strain-rate field over the plate boundary decadal timescales. A deployment strategy must cover the plate boundary zone while at the same time recovering the spatial variations in strain rate. At present, GPS measurements are the best method for providing the long-term stability needed to define the deformation strain field on timescales of decades, although advances in InSAR technology may make this feasible, at least in some environments. To recover both this field and the important superposed temporal variations, we propose a multi-tiered approach. First, we plan to deploy a 'backbone' network of CGPS receivers that will span the entire plate boundary zone at 100-200 km spacing and will consist of about 100 GPS receivers (**Fig. 8**). This backbone will provide a synoptic, long-wavelength map of deformation throughout the plate boundary zone. It will also constitute a strong fiducial network for firmly tying in new measurements that will continually refine this first-order picture and permit selective densification in some regions. The strain rate field, however, is strongly heterogeneous, and it is necessary to resolve this localized deformation. The backbone will be inadequate for this purpose. Measurements to date show that deformation zones are tens of kilometers to a few hundred kilometers wide, with large intervening blocks that are inactive or much more slowly deforming. We will instrument these high-strain-rate regions in two ways. Highest priority zones will be covered by CGPS clusters optimized to recover the localized strain.

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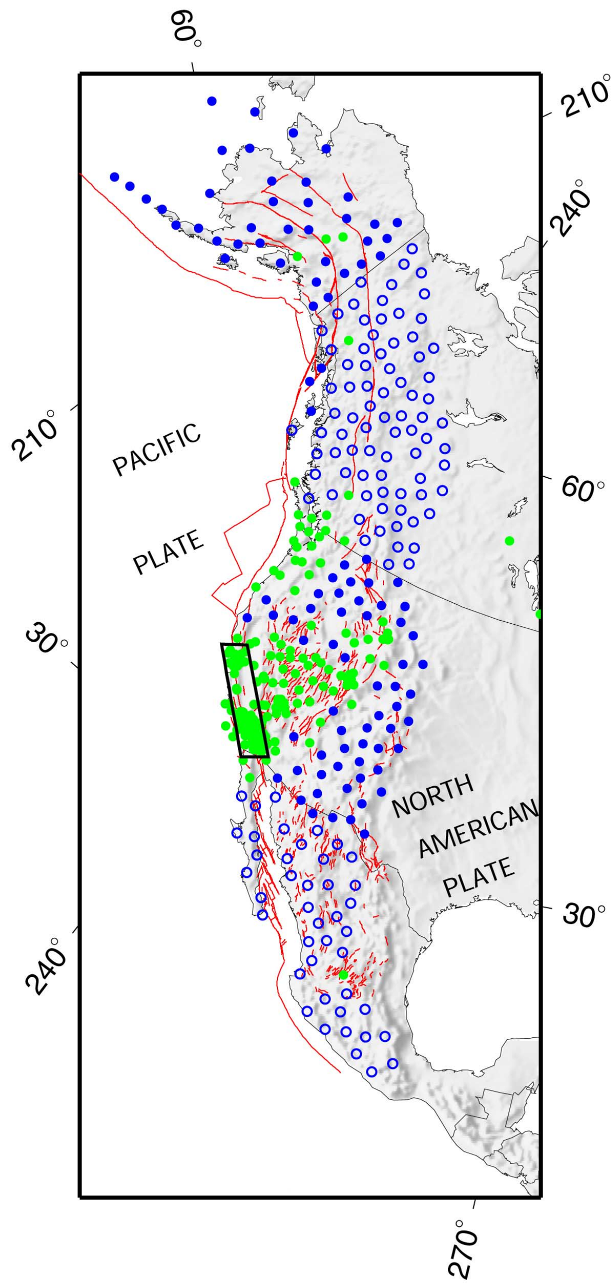
1. SCIGN: Southern California Integrated Geodetic Network, BARD: Bay Area Regional Deformation Network. BARGEN: Basin and Range Geodetic Network, EBRY: Eastern Basin and Range Network (U of Utah), and PANGA: Pacific-Northwest Geodetic Array.

These areas include, of course, the San Andreas Fault system, but also other areas within Cascadia, Alaska, and the Basin and Range. There will be other high-strain areas that cannot be instrumented with CGPS, due to limited resources, and for these we will utilize survey-mode GPS (SGPS). A dedicated pool of 100 new GPS receivers is requested for this purpose. The SGPS work would be carried out by individual P.I.s as part of programmatic research, but the PBO would supply the receivers. Historical seismicity, existing deformation data, as well as simulations based on known active fault distributions, expected internal stresses, and plate boundary driving forces will be used to guide CGPS and SGPS network design and deployment. The decadal deformation field is also important for measuring interseismic strain accumulation for the **earthquake cycle**. Decades of experience with existing networks in California provide guidelines for the required station spacing. The plate boundary zone across the San Andreas system is on the order of 100 km wide and contains from one to three primary active faults, with numerous lower-slip-rate faults.



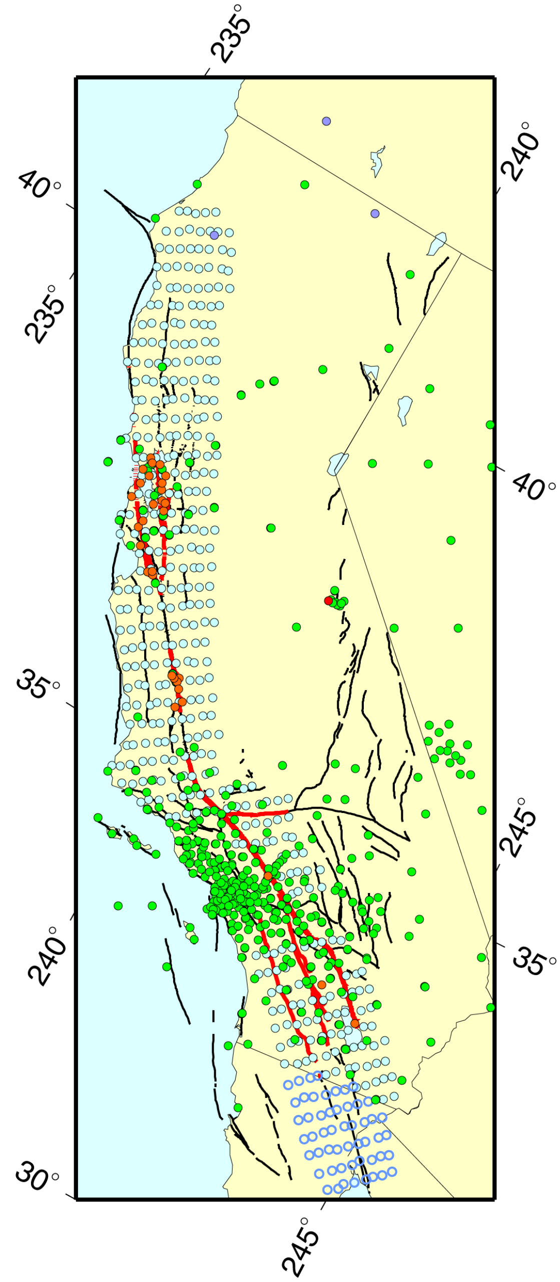


Figure 8a: (left) Proposed backbone array of continuous GPS receivers to capture the long-wavelength decadal field. Instrument spacing varies between 100 and 200 km. Filled blue circles give presently planned locations of stations in the United States, open blue circles in Canada and Mexico. Existing sites are shown in green.

Figure 8b: (right) Possible distribution of CGPS and strainmeters for an instrument cluster to cover the San Andreas Fault system, consisting of 400 new CGPS receivers and 175 new strainmeters. Filled blue circles give presently planned locations of GPS stations in United States, open blue circles in Mexico. Existing sites shown in green. Existing strainmeters shown in red. New strainmeters would be deployed along the most seismogenic portions of San Andreas fault system (highlighted in red) according to scheme shown in Figure 8c.

To accurately resolve the deformation associated with the various faults requires on average about 10 GPS sites per transect across each deformation zone. Transects should be spaced at something like the interseismic fault-locking depth of 10 to 15 km. This rule of thumb suggests that at least

100 GPS sites per 100-km-length of deformation zone are required. Given the roughly 1000-km length of the San Andreas Fault system, about 1000 GPS receivers are needed. We propose to accomplish most of this objective by deploying 700 CGPS receivers (400 new sites) and to cover the remainder of the fault system with SGPS sites **(Fig. 8b)**. This combination provides us with excellent spatial resolution, and with the opportunity to observe time dependence (see below).

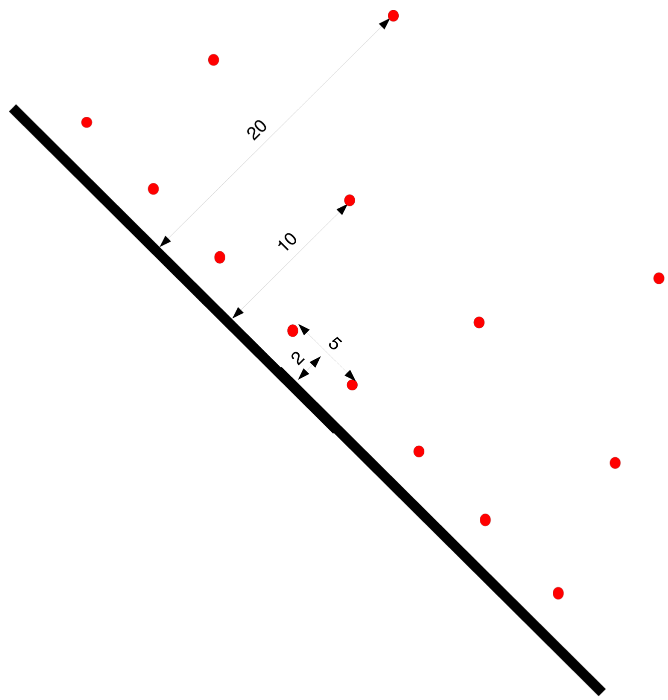


Figure 8c: Strainmeters would be deployed along active faults in the configuration shown to optimize both the detection of transients (stations close to fault) and resolution of location (those farther away). Distances in kilometers.

Much of the relevant plate boundary zone is offshore and is therefore inaccessible to land based measurements. New technology for seafloor geodetic and strain measurements for the first time make it possible to monitor deformation on the sea floor. While these methods are still somewhat experimental, they have the possibility of revolutionizing our understanding of subduction zone processes and of better constraining the offshore motion along the San Andreas Fault system (see Appendix A).

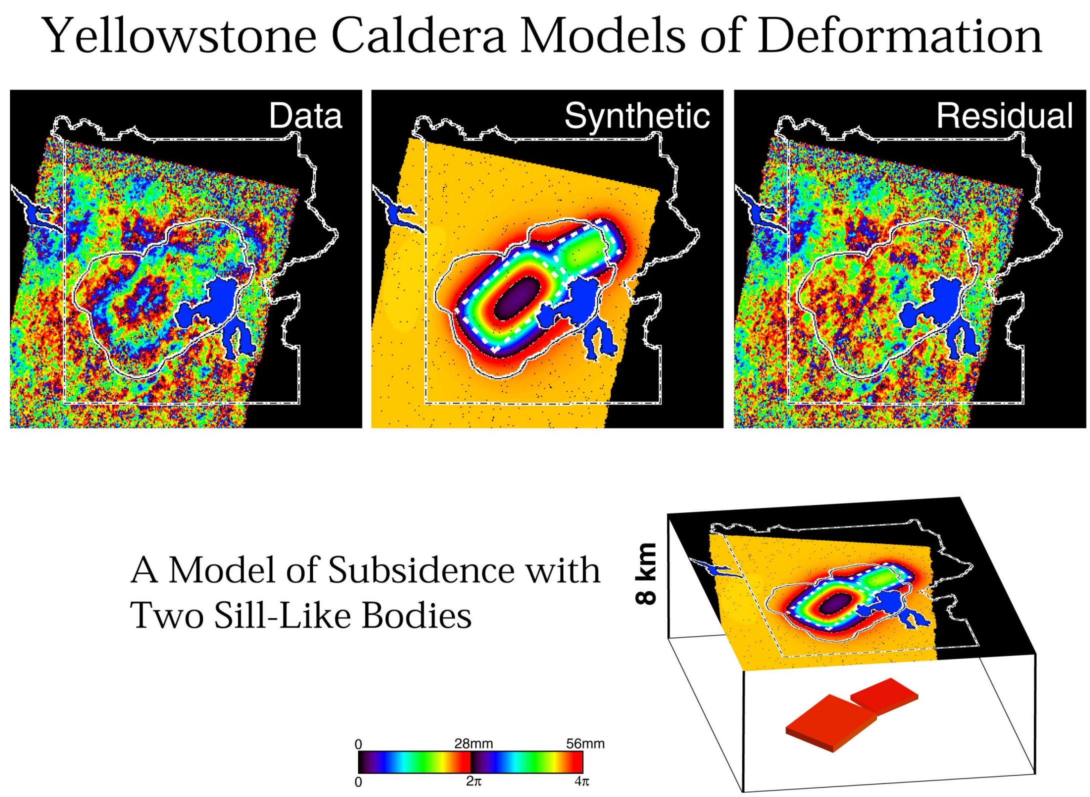


Figure 8d: Schematic view showing model sills beneath Yellowstone caldera and synthetic interferogram generated from this model. InSAR would be used in the design phase to guide the placement of both GPS receivers and straimeters around volcanic areas.

3.2.2 The Time-Dependent Deformation Field

Temporal variations from the decadal deformation field are most relevant for **earthquake** and **volcanic** processes. Observing these variations requires the close integration of strainmeters, CGPS, and InSAR. Simulation experiments will be carried out to design optimum arrays to maximize detection capabilities for the observation of a variety of transient phenomena. For example, the time-dependent inversion method of Segall and Matthews (1997) can be used to determine the capability of GPS and strainmeter networks to image transient fault-slip processes. The method is based on the fact that slip on faults produces spatially coherent signals across an array of instruments, whereas local noise processes are spatially incoherent. Preliminary simulations suggest that use of array processing techniques decreases the detection threshold for preseismic fault slip by at least an order of magnitude over what can be achieved from individual strainmeters. Such simulations demonstrate how proposed field deployments can use instrumental performance parameters (sensitivity, noise characteristics) to derive an optimum mix and spatial distribution of instruments to detect fault slip with maximum resolution and minimum uncertainty. They can be viewed as the Earth Science analogue of designing the operating characteristics of a telescope, in this case one that is downward looking. Extensive previous strainmeter experience in California indicates that we require three to four strainmeters at different distances from the principal faults, spaced at the fault locking depth. This estimate suggests that 20-30 strainmeters per 100 km of fault zone length are required for adequate monitoring (**Fig. 8c**). To detect longer period transient motions, a complementary network of at least 100 permanent GPS sites per 100-km-length of fault zone is also required. To maximize event detection these dense networks will be deployed on the San Andreas system where slip rates are high and instrumentation can be located closest to points of earthquake nucleation and aseismic slip.

Post-seismic transient deformation provides valuable constraints on post-seismic fault slip and fault zone constitutive behavior, as well as crust and upper mantle **rheology** in actively deforming regions. InSAR methods can presently map the spatial distribution over ~35 day orbital cycles, providing strong constraints on the geometry and locations of sources of deformation at depth and coarse resolution of their time behavior in the months to years following the earthquake. InSAR is also ideally suited to provide the 'initial conditions' for postseismic deformation, the coseismic deformation and fault slip distribution that drives subsequent stress relaxation processes. Thus, dense spatial and temporal resolution is necessary to make major breakthroughs in defining post-seismic fault behavior, lithospheric relaxation processes, and constraining controlling parameters. To accomplish these objectives InSAR coverage over the entire plate boundary zone is needed. In addition, immediate post-seismic deployment of dense CGPS arrays around the earthquake rupture zone will be required. The required number of CGPS sites would depend on earthquake size, but 50 sites per 50 km of fault would be needed. Some or all of these GPS receivers could be supplied immediately by loans from the 100 SGPS instrument pool and reimbursed later.

In the case of observing **magmatic processes**, locating magmas and following their temporal evolution requires InSAR mapping and deployments of both CGPS and strainmeter networks. The design phase would be guided by InSAR imaging to identify the locations and geometry of the sources of deformation. **Figure 8d** shows an example from the Yellowstone caldera, where InSAR methods have been successfully applied to identify two 8-km-deep sill-like bodies occupying most of the caldera floor and detect fluid migration between them (Wicks et al., 1999). With sources identified in this manner, simulations for detecting fault slip can be applied to optimize CGPS and strainmeter nets for maximum detection sensitivity and tracking of magma movement in time and space. Design phase modeling will clarify the optimal configuration, but previous experience on well-instrumented volcanoes in Hawaii and Long Valley suggest that networks would consist of 15-20 CGPS and 4 strainmeter sites. Surface strain and displacement measurements cannot necessarily discriminate between hydrothermal and magmatic processes. Since water and magma have dramatically different densities, they will have easily distinguished gravity signatures. For this reason precise gravity measurements, both relative and absolute, are a worthwhile addition in the study of volcanic systems.

As part of the basic PBO configuration, we propose to deploy permanent instrumentation, CGPS plus strainmeters, around six currently actively deforming magmatic areas. These are the Yellowstone hotspot, Long Valley, two from the Cascades, and two in Alaska and the Aleutians. The criteria are that some of these regions must be active now, they must sample a range of magmatic behavior, and they must be relevant for the problem of volcanic hazards. Because the Yellowstone hotspot is a much larger magmatic center than the other identified targets, larger networks will be needed for adequate monitoring.

3.2.3 The Long-Term Deformation Field

The long-term (greater than 100 years) surface deformation field, supplied by geological measurement, is a basic observational component of PBO that extends the geodetically-determined time-history of plate boundary deformation back through the Holocene and beyond. This element provides crucial information for the space-time distribution of **earthquakes** and

magmatic activity, as well as constraining the general **evolution** of tectonic deformation through comparison with the decadal field obtained from geodesy. The key to fully exploiting the geologically determined deformation field is in closely coupling it to the geodetically determined field. Indeed, the combined geodetic-geologic time series from PBO will be the first comprehensive dataset for a major plate boundary deformation zone, providing an unprecedented basis for understanding both the western North American plate boundary system as a whole, and the general problem of the dynamics of tectonic and magmatic systems. By coupling the GPS deployments with an appropriately coordinated program of **paleoseismology**, **volcanology**, and **tectonic geomorphology**, PBO will increase by two orders of magnitude the number of comparable geologic/geodetic measurements, and integrate them into an unprecedented image of the history of activity across an entire plate boundary deformation zone. Areas of focus for obtaining the long-term deformation field will thus be closely coordinated with the geodetic deployments, such as along the San Andreas Fault system. In particular, efforts will be concentrated in areas of cluster deployments of CGPS and SGPS instrumentation.

Retrieving the long-term field requires three specific classes of information (see Appendix B): (1) remote sensing data, which establish a template for (2) field-based displacement measurements, providing the spatial constraint on kinematics, and (3) laboratory-based geochronologic measurements, providing the temporal constraint. High-resolution topographic and tonal (color and/or gray shading) data are needed along all active plate boundary deformation zones and eruptive centers to identify deformational structures and to perform measurements such as fault offsets. These data will include either radar-based (e.g., GeoSAR) or laser-based (e.g., Airborne Laser Swath Mapping, ALSM) topographic imaging with vertical posting in the 10-30 cm range, and high-resolution digital imaging in the visible range at low sun-angle with pixel size in the 10-20 cm range.

Field measurements of displacement will then be made of the surface and shallow subsurface along and near active deformation zones and eruptive centers, motivated and guided throughout by the topographic and tonal database. Access to the shallow subsurface will require, depending on the details of the feature under investigation, surface mapping, trenching, shallow (~10 m) boreholes, ground-penetrating radar, and shallow seismic reflection profiling using low-energy sources. The actual measurements, especially surface mapping and trench logging, will be made using kinematic GPS, where one receiver is used as a stable reference and the other is used to measure relative displacements of offset features.

3.2.4 The Subsurface Deformation Field: The Role of Seismology

Since both **earthquakes** and **magmatic processes** originate in the crust, it is essential to probe the depth-dimension of deformation to understand them. To some extent this can be done through inversion of surface geodetic data. Yet, a direct constraint on subsurface strain is highly desirable, and this requires the use of **strain indicators**. There are several ways of approaching this problem. The most straightforward is the monitoring of microearthquake activity since it, by definition, represents the radiated component of subsurface deformation. It is essential to reduce the magnitude threshold of events, because the very smallest events most likely correspond most closely to slow aseismic deformation (in the normal sense). These considerations necessitate the use of ultrasensitive borehole seismometers, which serve to reduce dramatically the noise level. For this reason we have included with each borehole strainmeter a borehole seismometer for the continuous monitoring of microearthquake activity.

These small events can be used to map detailed structure using relative relocations at tens of meters resolution, as well as temporal variations in seismicity (e.g., Nadeau and McEvilly, 1997,

1999) and crustal elastic properties (e. g., Karageorgi et al., 1997). There is an excellent example of a microearthquake strain transient from Parkfield. In addition to the long-term transient seen in surface deformation measurements (Gwyther et al., 1996, Gao et al., 2000) mentioned earlier, an analysis of the 11-year high-resolution microearthquake record at Parkfield reveals systematic spatial and temporal changes in the San Andreas Fault slip rate indicated by the repeat times of identical multiple events that were synchronous with these other indicators of deformation. For the 2.5-year period beginning in October 1992, the analysis defines a deformation pulse of increased slip leading to and accompanying the series of M=4.6 – 5 earthquakes that occurred during that time (Nadeau and McEvilly, 1999). Borehole seismometer data from sites along the northern Hayward fault are currently being analyzed using this approach to investigate the detailed fault structure and to gather slip-rate information at depth (McEvilly et al., personal communication). The actual relationship between microearthquake activity and strain is important to determine. This objective can be accomplished by drilling into a seismically active area, as is proposed by **EarthScope** initiative **SAFOD** (see Section 7.2).

Seismic activity is also a means of providing a constraint on strain through estimates of source mechanisms and, with certain assumptions, their interpretation in terms of stress orientation. This methodology also provides point-specific estimates of stress on individual faults that can be directly compared to surface observations. Data from existing regional (fault and volcano specific) and national seismic networks can be used for this purpose (see discussion of **EarthScope** initiative and **ANSS** in Section 7.4).

Another approach to constraining subsurface deformation is through the use of seismic tomography. It is possible, for example, to image fault zones and magma chambers since they have distinctly different seismic velocities. The internal structure of faults, which may be replete with fluids, cracks, and pulverized rock, are most clearly observed by the use of fault-zone guided waves. These waves, observed for fault-parallel propagation paths, constitute a new tool that is being explored to reveal the important details of fault structure. In addition, it is sometimes possible to detect temporal variations in fault structure, as has been observed following the 1992 Landers earthquake (Li et al., 1998). Faults can also be imaged through their influence on horizontal seismic reflectors. A fault-related vertical offset in the Moho, for example, would constitute evidence that the fault extends below the seismogenic zone and into the mantle. Both imaging techniques can make use of the portable instrumentation provided by **USArray**, another element of the **EarthScope** initiative (see Section 7.1).

3.3 Recommended Geodetic Deployment Configuration

The core equipment request consists of two arrays of geodetic instruments, GPS receivers and strainmeters, so as to have equal strain-rate sensitivity from the seismic band out to several decades (**Fig. 8).** The deployment strategy will have a two-tiered structure dictated by the scientific problems addressed. There will be a backbone deployment of continuous GPS receivers that covers the entire North American plate boundary zone from Alaska to Mexico, and from the west coast to the eastern edge of the cordillera, at 100-200 km spacing (Appendix C). This backbone will consist of approximately 100 new GPS receivers. Imbedded within the backbone array will be several clusters of instruments for problems related to the study of earthquakes and magmatic processes that require higher spatial and temporal resolution. While there are several candidates for earthquake-related deployments, there is unanimous agreement that one of the cluster areas must be the San Andreas Fault system (SAFS). There has already been significant progress toward such a deployment, in that there are 300 CGPS receivers operating (or already funded) along the San Andreas Fault system, 250 in the south and 50 in the north. On the basis of deployment criteria discussed above, an additional 400 receivers are needed: 75 in the south and 325 in the north and central regions. In contrast, it is anticipated that there will only be 45 strainmeters along the SAFS by 2002, and this number needs to be increased to about 220 (175 new strainmeters) to obtain adequate coverage and a fully integrated system throughout the entire SAFS. With such an integrated deployment, and given seismicity rates, we would expect, over a ten-year period, to ‘capture’ about 3, 30, and 300 events with magnitudes equal to or larger than 6, 5 and 4, respectively. We thus expect to record some large damaging events, but will also be able to make extensive use of and learn from the smaller, more frequent events that will be observed over this time. In addition, the PBO will consist of clusters around six magmatic centers. The present consensus is that these six centers would be Yellowstone, Long Valley, Mount Rainier, one other Cascade volcano, and two Alaskan volcanoes to be determined. With the exception of Yellowstone, each magmatic center will be instrumented with 15-20 CGPS receivers and 4 strainmeters. Yellowstone, because of its exceptional size, will require roughly three times the number of instruments. On the other hand, Long Valley presently or will shortly possess nearly sufficient instrumentation for the PBO, so that the total new instrumentation needed to address the magmatic objectives is 100 CGPS receivers and 25 strainmeters. The total complement of new instruments for this configuration is thus 600 CGPS receivers and 200 strainmeters. We note that each strainmeter installation will also include a borehole seismometer and tiltmeters.

There are several other promising areas for cluster arrays. These include the Queen Charlotte-Fairweather fault (Alaska), the Alaska-Aleutian subduction system, including the postseismic response to the 1964 Alaska earthquake, the Cascadia subduction system, the Mendocino Triple junction, and the Walker Lane (Basin and Range). For these areas we propose that the actual locations of PBO deployments be determined competitively based on scientific merit, in a manner yet to be determined. This leaves 275 GPS receivers for these other clusters, or some combination of GPS and strainmeters if that is preferred. We also require 100 additional GPS receivers for survey-mode GPS for densifying areas not sufficiently covered by CGPS, and for rapid response capability.

Finally, it is clear that the North American plate boundary zone passes through both Canada and Mexico, as well as the U.S. The problems that we seek to study do not stop at our borders. The San Andreas Fault system extends into Mexico. The cordilleran structures seen both in Alasaka and in the Pacific northwest of the U.S. extend without interruption into Canada. Thus it is essential that the U.S. collaborates closely with its Canadian and Mexican counterparts for a fully integrated Plate Boundary Observatory. While we do not seek funds for instrumentation to be deployed in these countries, contacts have been made with Mexican and Canadian counterparts, and there is keen interest by scientists in these countries to join in a collaborative effort (see Appendix D).

4.0 PBO Management

A strong management plan is critical for the success of an endeavor as large as the PBO, which will install a heterogeneous set of over 1000 new geodetic instruments over a vast area of western North America over a period of 4-5 years, and operate them for roughly 20 years. To put this into perspective, the largest single geodetic array in the United States, SCIGN in southern California, consists of 250 sites (about 150 sites now installed). With PBO sites being installed over an area several hundred times as large as southern California, the challenge of selecting sites, installing instruments, and operating them is substantial. Even the GEONET network in Japan, while of similar size in terms of number of instruments, covers an area only a fraction the size of the PBO. Neither of the projects mentioned above include the variety of instrumentation involved in the PBO.

A crucial management requirement is the need to enfranchise the broad range of scientists interested in using geodetic data to study the plate boundary zone. These include researchers who operate strainmeter and continuous GPS networks in the area, those who conduct survey (campaign) mode GPS in the area, researchers interested in using the GPS or strainmeter data, and researchers who use the resulting velocity or strain fields to develop and test models. Participation and representation from all of these groups will be needed to ensure that PBO is installed, achieves maximum scientific return, and is fully supported by the scientific community.

PBO should be run as a facility much like IRIS, the successful consortium for seismic instrumentation. The management structure would combine aspects of a community-driven, consensus-building grass-roots effort with a strong centralized management structure, representing and empowered by the scientific community. We expect that PBO management would reflect the needs of the broad scientific community, coordinate in a coherent way a far-flung and heterogeneous set of instruments, and maintain uniform standards over the array to the extent possible while allowing flexibility for local conditions. It should take advantage of local knowledge and expertise without over-burdening individual scientists or losing overall coordination. The PBO management should furthermore formulate priorities and adopt an overall network design consistent with the decisions of a broad community, and coordinate installation and operations activities such as locating and permitting sites, selecting and installing the proper mix of PBO instruments, operating them, and collecting, archiving, analyzing, and disseminating large quantities of data. PBO management must make major decisions on behalf of the community, coordinate support from a variety of agencies, perform financial administration, report back to funding agencies, award contracts, and oversee routine operations.

To accomplish these aims, we recommend that PBO management (shown schematically in **Fig. 9**) be directed by a Steering Committee. As with other facility organizations, the committee would be elected by representatives of PBO member institutions who would reflect the interests of the broad community of scientists interested in PBO-related research. The committee would receive input from working groups set up to address the variety of issues involving site selection, operations, and specific technical, regional, and scientific issues.

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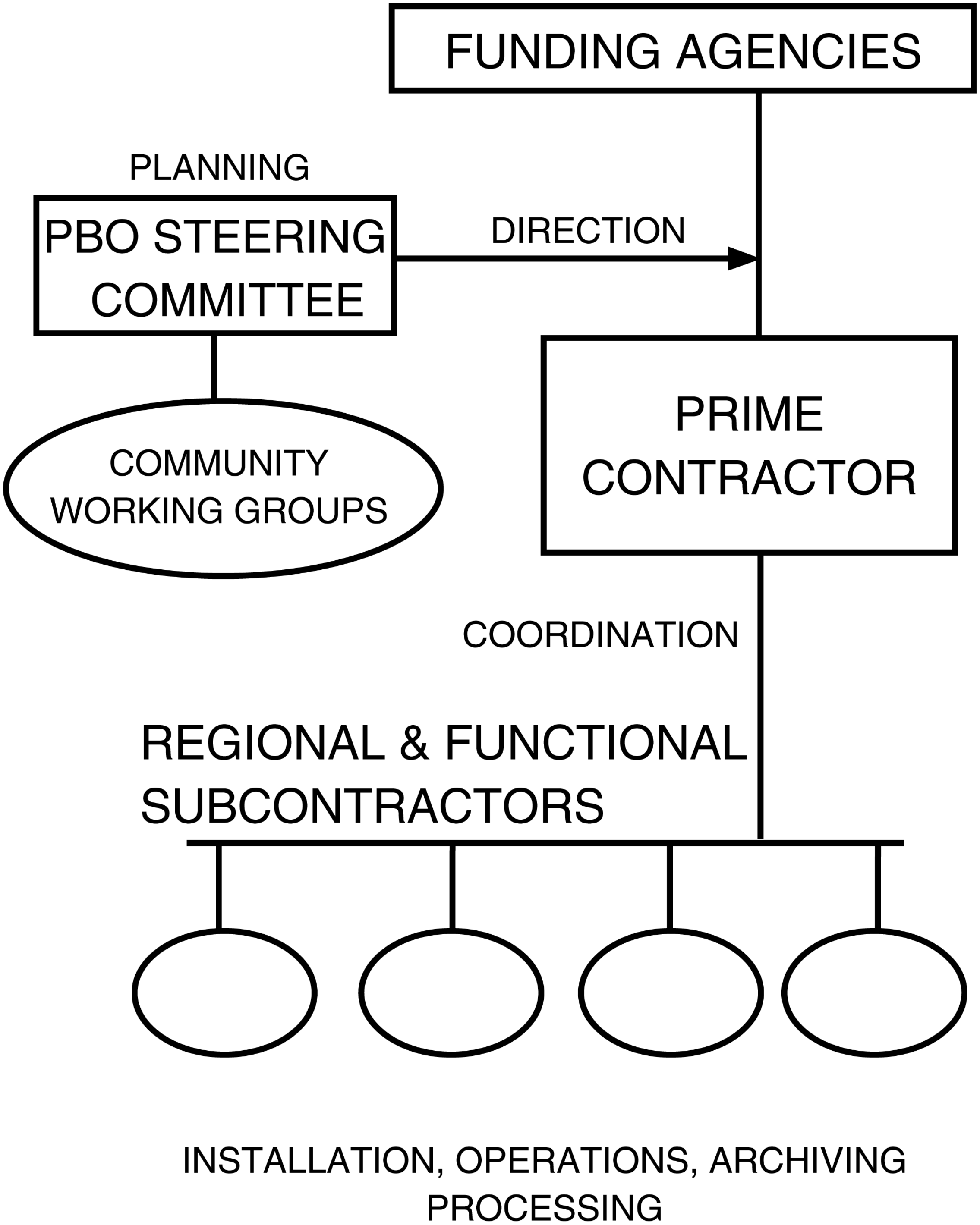

Figure 9: Proposed management plan for PBO.

The Steering Committee would direct the operations of a prime contractor. The role of the prime contractor would be to receive NSF funds and be responsible for carrying out PBO operations, in many cases via subcontracts to either regional networks (especially for the GPS component) or functional contractors (for example, it may make sense for a large fraction of the strainmeters to be operated by a single subcontractor). A natural choice for prime contractor would be UNAVCO, the national university consortium funded by NSF and NASA to support GPS networks and operations. Another alternative for running the PBO facility could be based on the operation of the successful SCIGN network.

In either case, PBO can build on existing research efforts within western North America. The most advanced is the continuous GPS component, with nearly 250 sites installed in western North America and another 150 planned. Existing CGPS arrays in western North America include SCIGN (southern California), BARD (northern California), BARGEN (Basin and Range), EBRY (eastern Basin and Range and Yellowstone), PANGA (Pacific Northwest), WCDA (western Canada), AKDA (Alaska), and an array in Mexico operated by Mexican institutions. These permanent GPS sites were installed to study specific scientific problems, many of which are included in the PBO scientific targets. We expect PBO to either cooperate closely with the current operating agencies or, where appropriate, assume responsibility for the operation (possibly via a subcontract to the current operating agency) of some or all of the sites of these networks. In many cases, these sites will have accomplished their initial scientific objective but would remain valuable parts of the PBO backbone array or instrument clusters. Another effort related to the PBO is a project funded by NSF which includes scientists from UC Berkeley, Carnegie Institution of Washington, Scripps Institution of Oceanography, and USGS to install an integrated array of borehole strainmeters, borehole seismometers, and continuous GPS sites in northern California, and operate a SAR downlink facility in southern California. Finally, existing strainmeter efforts include the Piñon Flat Observatory group, which has installed three long-base strainmeters in southern California, and the USGS group, which has installed more than 20 borehole strainmeters in California

5.0 PBO Implementation

Under the direction of the Steering Committee, the prime contractor will organize installation and operation of the PBO. The project may be divided into three main phases: design, installation, and operation. The design phase consists of determining the scope of the project, a process that is already underway. This phase includes not only decisions about the regional scope of the PBO, but also the prioritization of the specific scientific targets to be addressed through the PBO backbone network and instrument clusters and the determination of the optimal network design to achieve those goals. The design phase will also include detailed analysis and modeling to ensure that the final network design will meet the scientific goals of the project. The final part of the design stage, not yet begun, will be site selection, in which the general locations of all sites are specified. We recommend that the design phase include a series of workshops, supplemented by focused research efforts and in some cases reconnaissance measurements.

The installation phase begins with detailed site selection and permitting. This step is then followed by the actual site installation and testing before sites become operational. The process of obtaining permits for permanent geodetic installations can be slow and difficult, and in most regions is expected to be the limiting factor on the rate of site installation. Thus it is critical that at least some sites be selected well in advance of the first installations, and that appropriate personnel be allocated for both selection and permitting. Local experience and involvement is expected to be critical in this phase, both to avoid pitfalls from particular local conditions (such as security considerations, local noise sources, extremes of heat or cold, heavy snow accumulation, and limitations of regional infrastructure). We expect that the installation phase will be carried out by one or more installation contractors, with oversight and direction from the PBO Steering Committee via the prime contractor. Because of the specialized technical knowledge required by the various instrument types, it is likely that subcontractors will be used to install the different instruments. Installation for each instrument type could be done either by a single subcontractor with regional offices, or by several subcontractors each with a specific regional focus. A regional focus, either through regional offices of a single subcontractor or separate subcontractors, makes sense from a logistical point of view because in areas where the density of sites is high the installation can be done much more efficiently by a group stationed in that area. Since several alternatives are possible, it is likely that different approaches will be used in different areas.

The operational phase consists of routine operation, troubleshooting, archiving, and data analysis. These tasks will be coordinated by the prime contractor under Steering Committee direction. As in the installation phase, it may make sense to have different subcontractors handle the data from the different instrument types, although in some places it will be more efficient for all data from the same area to flow through one operating center. The fields of data archiving and distribution are evolving quite rapidly, so several models are worth considering. For example, the data could come directly to an archive center, as is done presently by UNAVCO and IRIS, or much of the data could first come to regional data centers. Most likely, a hybrid mode would evolve. The prime contractor might archive much of the data, but after the data have first passed through several operational, analysis, and archiving centers, following the model of the International GPS Service (IGS). The centers could be chosen from a host of universities, research laboratories, and government agencies. Ideally, we would achieve the benefits both of integrated central management and decentralized operations by groups close to the issues of specific sites. Routine operation and troubleshooting could be distributed or centralized, depending on a variety of considerations, including logistics. In the distributed case, the prime contractor would direct operations and distribute resources for operation and maintenance of sites. Whichever mode is chosen, the prime contractor will ensure that both GPS data and data products (e.g., site motion vectors) from all sites are easily available to all users. Data access will be accomplished via a seamless interface being developed by a multi-institutional group organized by UNAVCO.

6.0 Budget

Fiscal Year	GPS(N)	Strainmeters(N)					
New 2002	150	20					
New 2003	225	30					
New 2004	200	50					
New 2005	200	50					
New 2006	200	50					
Total Sites	975	200					
Costs(K$) Install Period	**Instrumentation GPS**[1]	**Instrumentation Strain**[2]	**Installation**[3]	**Operation**[4]	**Airborne Imaging**[5]	**Science**[6]	**Total**
2002	3,250	1,000	3,000	-	3,000	1,000	11,250
2003	5,125	1,500	5,125	720	2,000	1,000	15,470
2004	5,000	2,500	6,250	1,950	-	2,000	17,700
2005	5,000	2,500	6,250	3,450	-	2,000	19,200
2006	5,000	2,500	6,250	4,950	-	3,000	21,700
5 Year Total	**23,375**	**10,000**	**26,875**	**11,070**	**5,000**	**9,000**	**85,320**
Next 5 Years							
2007				6,450		4,000	10,450
2008				6,450		4,000	10,450
2009				6,450		4,000	10,450
2010				6,450		4,000	10,450
2011				6,450		4,000	10,450
5 Year Total				**32,250**		**20,000**	**52,250**
GRAND TOTAL							**137,570**
Breakdown by Category							
Equip/Install (GPS)		45,250					
Equip/Install (Strain)		15,000					
Equip/Install (Total)		60,250					
Maintain Sites, 10 yrs.		43,320					
Airborne Imaging		5,000					
Science		29,000					
Total		**137,570**					

1. GPS instrumentation includes GPS receivers/antennas (12), data transmission equipment (5), meteorological station (2), and environmental security enclosure (6) for total of $25K/site.
2. Borehole strain/seismic instrumentation includes 3-component strain (20), seismometers (5), tiltmeters (5), pressure transducers (5), and data loggers (15) for total of $50K/site.
3. GPS installation is $25K/site including siting, permitting, and installation. Borehole strain/GPS installation (includes drilling/casing of 200-m boreholes and installation of borehole instrumentation) is $50K/site.
4. Operation and maintenance includes technical staff for field maintenance, data downloading, calibration, reduction, and archiving at $6K/site per year.
5. 25,000 km of 10 km aperture data at $200/km.
6. Science includes data analysis and interpretation, geological fieldwork and analyses, and workshops.

7.0 Relation to Other Elements of EarthScope

7.1 USArray

Seismic Imaging of the Subsurface

The Plate Boundary Observatory and USArray share the goal of illuminating and understanding the faults in western North America. Seismic and geodetic measurements have provided most of the constraints on fault mechanics so far, and are likely to continue to do so in the future. USArray can aid PBO in studying earthquakes on the San Andreas and other fault systems, as well as in mapping the tectonic structures.

Permanent and temporary USArray deployments will provide numerous hypocentral locations and focal mechanisms of small earthquakes, sometimes in areas that have not previously had a dense seismic network in the neighborhood. Maps of seismicity and focal mechanisms can be routinely interpreted to infer the style and geometry of tectonic deformation. Additional parameters such as earthquake stress drop and orientation and extent of the rupture plane are also useful for understanding the dynamics of the fault system.

The many seismometers and the flexibility of the USArray deployment strategy will enable innovative experiments to image faults. Recent work shows that active faults are sometimes marked by narrow zones of reduced velocity, which may focus guided waves, and that fault zones sometimes reflect seismic waves, opening the possibility of mapping the fault plane in the same way that exploration geophysicists map oil fields. The ability to directly map the features of fault planes would improve significantly our ability to infer the location of asperities and barriers in future, as well as past, earthquakes.

Coordinating the Geological Components of USArray, PBO

As elements of EarthScope, USArray and the Plate Boundary Observatory (PBO) share both scientific goals and implementation needs. As one of its scientific goals, PBO will measure on-going deformation of the plate boundary. This motion manifests itself on the short timescale by earthquakes and on the longer timescale by the development of deformation structures in the crust such as faults and folds. While deployed in the western U.S., the transportable seismometers of USArray will provide a sensitive monitor of seismicity, tracking earthquake activity over the entire western U.S. In addition, the imaging capabilities of USArray, both seismic and magnetotelluric, will help identify subsurface structures, such as buried faults and major structural breaks in the crust that can be both the expression of long-term deformation and the locus of modern strain relief on the plate boundary. Similarly, the high-resolution topographic imaging provided by the surface geology component of PBO can be overlain directly on the subsurface images from USArray for detailed investigation of the connection between surface topography and underlying geologic structures. In addition, in areas where the longer-term behavior of faults is studied by the paleoseismology efforts of PBO, the flexible pool of USArray seismometers could be used to extend these studies to deeper levels in the crust to understand the full extent of the crust disrupted by faulting. Similarly, in studies of the time history of particular volcanic centers to be carried out as part of PBO, the flexible component of USArray could be used to provide detailed subsurface images beneath the volcano, perhaps even tracking the magmatic plumbing system that feeds surface eruptions.

The surface geology component of PBO extends the analysis of the dynamic behavior of the plate margin to the thousand-to-million year timescale. This timescale is critical to the characterization of episodic behavior of both faults and volcanic systems that will help interpret, and perhaps predict, current behavior. USArray extends this timescale further, to millions or billions of years, to investigate both how the longer-term activity on the plate boundary has shaped the geologic structure of the crust and, in turn, how the geologic history of the crust determines the way in which strain release is partitioned across the plate boundary. Both of these approaches rely heavily on modern geochronologic methods and will require an investment in these laboratory-based techniques as outlined in both the PBO and USArray white papers.

Technical/Operational Cooperation

PBO and USArray also share some overlapping technical and operational aspects where both projects benefit by cooperation in design and implementation. USArray will require the construction of seismometer vaults throughout the western U.S. Where possible and mutually beneficial, these installations could be upgraded to the high-stability geodetic monuments needed for the permanent GPS network of PBO thereby serving both programs. Co-location of sites also could allow shared telemetering and power sources, thereby allowing a distribution of the associated costs between USArray and PBO. In those areas where co-location is not possible, high-quality GPS documentation of each USArray site during its installation would allow these sites to be revisited later for campaign-style measurements, thereby increasing the density of GPS coverage of PBO. Critical to both PBO and USArray objectives is the measurement of geologic time. Both PBO and USArray will rely on the same infrastructure for these laboratory-based geochronological techniques. The number of geochronologic measurements needed to meet the combined goals of USArray and PBO may well exceed the capability of existing laboratories, necessitating augmentation of these facilities or the creation of new laboratories for geochronology.

Education and Outreach

A critical component of both PBO and USArray will be effective public information and outreach. Both projects, as integral components of EarthScope, present tremendous opportunities for increasing the public awareness of earthquakes, volcanic eruptions, and Earth science in general. Both PBO and USArray are proposing significant education and outreach programs; these programs share some common aspects but also complement each other. The proposed PBO outreach program, as outlined in more detail in Appendix E, will focus on public awareness of earthquakes, volcanic eruptions and their risk to society throughout the western United States. The USArray education and outreach program will be of national extent, reflecting the eventual national coverage of USArray. When the transportable seismic array is located in the western United States, the PBO and USArray programs will complement each other in content, with both programs contributing to general K-12 education programs, and with PBO additionally including education and outreach to city planners, developers, and other agencies specifically concerned with natural hazards.

7.2 SAFOD

The establishment at Parkfield, California, of the San Andreas Fault Observatory at Depth (SAFOD) will provide direct observation at seismogenic depths of stress, physical state, and seismicity at a site along the San Andreas Fault system, where the probability of occurrence of a magnitude 6 earthquake is relatively high. Thus, SAFOD will provide a unique calibration experiment, permitting comparison of surface deformation measurements with observations being made on the San Andreas Fault at 3-4 km depth. The results will have strong application to many other PBO deployments along the San Andreas Fault System where at-depth observations cannot be made, significantly amplifying the value of these measurements to the mutual benefit of PBO and SAFOD.

7.3 InSAR

InSAR represents a highly complementary method of observing plate boundary deformation, as mentioned earlier. While GPS most accurately measures horizontal displacement, InSAR is more sensitive to vertical displacement. GPS yields highly accurate point measurements of displacement, whereas InSAR provides coherent coverage over broad regions. The integration of GPS and InSAR thus constitutes a particularly powerful space-based observational system to study the plate boundary zone. It has been especially useful in applications to major crustal deformation events such as earthquakes and volcanic eruptions, as well as regional ground failure and sympathetic faulting. It is clear that InSAR will play a significant role in the measurement of plate boundary deformation and in the observational capability of the Plate Boundary Observatory.

7.4 ANSS

Permanent seismic monitoring systems such as the Advanced National Seismic System (ANSS) provide data that are used in a range of activities that include emergency response, seismic hazards assessment, earthquake engineering, basic and applied earth science research, public information, and education. The ANSS directly benefits and complements the Plate Boundary Observatory by providing long-term monitoring of seismically active regions of the western United States, with the shared goal of understanding the physics of earthquake occurrence and mapping and illuminating active faults in the western United States. Improved standardization of recording and data distribution between permanent seismic networks results in higher quality earthquake catalogs and waveform datasets that can be used in a variety of PBO research activities. For example, much of the western United States is covered by permanent seismic networks and steadily improving instrumentation. Data from individual or integrated seismic networks in the region can contribute directly to better imaging of crust and mantle structure that are key to understanding the geologic development of the plate boundary zone on a variety of scales. Since the ANSS primarily involves permanent real-time seismic monitoring, it has an established telecommunication infrastructure that can be used to telemeter other data streams from PBO stations, such as GPS sites. Combined, PBO and ANSS observations provide the spatial density and spectral content necessary to significantly improve the understanding and mitigation of earthquake and volcanic hazards, as well as our understanding of the geodynamic development of western North America.

8.0 Summary and Recommendations

We propose to deploy a Plate Boundary Observatory to address several fundamental scientific problems associated within the actively deforming region of western North America. The three broad problem areas are plate boundary dynamics and evolution, the physics of earthquakes, and magmatic processes. The core of the proposal is the deployment of geodetic instrumentation to provide an unprecedented increase in our ability to observe plate boundary deformation, in terms of broad coverage and temporal and spatial resolution. The temporal range of observation will be extended back in time through geologic observations, and to the subsurface by seismic imaging. We expect this facility to be utilized by a significant portion of the U. S. Earth Science community. Based on participation at the PBO workshop, present membership of the GPS consortium UNAVCO, and anticipated growth in this area of research, it is expected that about 150 scientists and as many students, from about 120 U.S. institutions would be involved in either the operations of the facility or analysis of the data. In addition, we expect great interest from foreign scientists, especially from countries within plate boundary zones, amounting to approximately half the level of U.S. scientific participation.

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Appendix A. Instrumentation

A. Introduction

The PBO is an integrated array of scientific instrumentation in western North America that will provide a comprehensive space-time image of plate boundary deformation on the broadest possible temporal (from seconds to decades) and spatial (from tens of meters to hundreds of kilometers) scales. Although the PBO is conceived as a single strain-measuring instrument/ facility covering western North America, no single technique spans this spatial and temporal range. Therefore, our goal is an optimal integration of several highly complementary techniques (**Fig. 5**): continuous GPS (CGPS), interferometric synthetic aperture radar (INSAR), strainmeters, borehole seismometers, and other instrumentation such as precise GPS/acoustics ("sea-floor positioning") and absolute gravimeters.

At the long-period or far-infraseismic (periods greater than one month) end of the geodetic deformation spectrum, CGPS is particularly suitable to study decadal motion and its spatial variations, as well as large regional transients, such as the post-seismic deformation from major earthquakes. Furthermore, monitoring of the plate boundary can be extended to the off-shore ocean bottom by combining CGPS with precise acoustic ranging to seafloor transponders. Strainmeters, in contrast, can sample local spatial scales and have excellent sensitivity (2 to 3 orders of magnitude greater than GPS) at periods in the short-period geodetic band (from minutes to a month). InSAR provides an excellent means of observing deformation over broad areas. It is also capable of tens of meters spatial resolution at monthly or greater intervals. InSAR has proven to be a powerful tool to characterize the large-scale coseismic deformation from earthquakes, and it can also resolve small-scale deformation features such as shallow creep associated with earthquakes and postseismic and possibly interseismic deformation. Strain rate sensitivity of INSAR, however, is somewhat lower than GPS.

Unlike GPS and InSAR measurements, which provide a strictly geometric determination of deformation, gravity measurements are sensitive to the local mass distribution and topography. A combination of both surface geodetic measurements and gravity can provide unique information into deformation dynamics. For example, gravity time series can monitor changes in the subsurface mass distribution and in combination with CGPS, surface vertical displacements. This kind of instrumentation is most relevant to volcano monitoring.

Seismic observations provide the most straightforward means for detecting deformation at seismogenic depths. In particular, microearthquake activity constitutes the radiated component of active deformation in fault zones and as such provides important information about fault zone structure, kinematics, and processes.

B. Proposed Instrumentation

1. Global Positioning Systems

GPS is now a well-established positioning tool for geophysical research. Changes in site positions over time are used to infer global plate motions, plate boundary deformation, and permanent deformation associated with earthquakes such as coseismic and postseismic motions. Although GPS field surveys provide useful information, the PBO facility will focus primarily on permanent observatory-type installations. Compared with field surveys, CGPS provides improved temporal resolution, enhanced accuracy and reliability, and the means to better characterize the errors in GPS position measurements and thereby obtain more realistic estimates of derived parameters such as site velocities.

Instrumentation is relatively inexpensive, and measurement accuracy is about 1-2 mm in horizontal position and 5-10 mm in vertical position. This precision can be obtained over short length scales for observation periods as short as several hours, and for length scales of several hundred of kilometers or more with daily averaged observations. Continuous GPS measurements can be used to estimate steady state velocities with a precision of 0.5-1.0 mm/yr or better with several years of data. To achieve these long-term accuracies it is often necessary to install extremely stable monuments, such as the deeply anchored monuments used in most high-precision networks today. The need for site stability complicates site selection and raises the cost of GPS installations.

Since the early 1990's, CGPS arrays have been established in Japan (~1000 stations today) and in the PBO focus area of western North America. Today, there are (or will be within a year) ~ 400 CGPS sites in the PBO area. The sites are organized in arrays, including SCIGN (southern and Baja California), BARD (northern California), BARGEN (north and south Basin and Range), EBRY (east Basin and Range and Yellowstone), PANGA (Pacific Northwest), WCDA (western Canada), AKDA (a fledgling effort in Alaska), and UNAM (an ongoing effort in Mexico) (**Fig. 8a**).

Centimeter-level motion of the Juan de Fuca plate (a plate entirely submarine) has recently been demonstrated with annual visits to a site 150 km offshore Vancouver Island. The positions of the sea floor transponders were measured relative to a sea-surface platform (buoy/ship) by acoustic ranging from the platform to the transponders. The sea-surface platform was then tied to the global frame with CGPS positioning of the platform relative to static stations ashore. Establishing such GPS/acoustic seafloor transponder sites at a few key locations within the PBO region can help build understanding of underlying physical processes, such as the updip extent of the locked portion of subduction thrust faults along the Cascadia subduction zone and the Aleutian Trench, the pattern of plate movement within the so-called Gorda Deformation zone around the Mendocino triple junction, and the width of the shear zone offshore of California. The expense of GPS/acoustic geodesy is essentially driven by the periodic requirement for ship time, while the cost of monumentation is no more expensive than on land. These costs could be significantly reduced by establishing permanently stationed buoys above the transponder arrays.

2. Strainmeters

The primary advantage of strainmeter instrumentation is its superior sensitivity to strain at periods ranging from the seismic band (periods of one hour or less) to about a month, as well as sensitivity to localized phenomena. Borehole strainmeters use a hydraulic sensing technique to achieve a volume strain sensitivity of 10^{-12} with constant frequency response from 0 to more than 10 Hz and a dynamic range of about 130 dB. Long-base laser strainmeters at the surface use optical interferometry to measure changes in distance and anchoring of the ends for stability. Actual signal-to-noise levels for both systems depend on the spectrum and magnitude of Earth noise. Taking noise into account such instruments are 2-3 orders of magnitude more sensitive than GPS at short periods (day to month); at longer periods the advantage declines significantly, though the noise level of long-base instruments appears to be comparable to CGPS.

Because of the difference in modes of measurement, and hence in potential noise sources, between the long-base and borehole sensors, it may be worthwhile to co-locate a few long-base instruments to provide a check on signals that might be observed by other instruments, although the high cost of the long-base instruments severely restricts the number of such comparisons.

Arrays of borehole strainmeters have been operating for a number of years near selected sections of the San Andreas Fault in California. A modification of Sacks-Evertson dilatometers to three tensorial components has been deployed since 1985, and there are presently at least a dozen operating in Japan and close to 20 dilatometers operating in California. Several three-component borehole strainmeters of the Gladwin type (Gladwin and Hart, 1985) have also been deployed in California. Long-base laser strainmeters are currently operating at Piñon Flat Observatory and Durmid Hill in southern California, with new sites under construction in the Los Angeles basin as part of the SCIGN project. Recently, NSF funded an MRI proposal to increase the density of borehole strainmeters and seismometers in the San Francisco Bay area, with instruments to be co-located with new CGPS sites and supplemented with INSAR imagery.

3. Remote Sensing

a. Space-based Platform (InSAR)

InSAR has proven to be a powerful tool to characterize not only the large-scale coseismic deformation from earthquakes and magmatic processes, but also small-scale deformation features such as shallow creep associated with earthquakes and postseismic and interseismic deformation. The latest dramatic example of the power of InSAR was demonstrated after the October 16, 1999, Hector Mine M_w 7.1 earthquake in southern California. InSAR with the current ERS satellites has a number of significant disadvantages, however. It provides only a one-dimensional range-change estimate, and without additional information it is not possible to discriminate vertical from horizontal displacements. Due to the ~23° off-vertical incidence angle of the ERS SAR, horizontal displacements do not map strongly into the range-change vector. Orbital errors appear in the form of broad range gradients across an interferogram, which are difficult to separate from regional strain gradients. Atmospheric delays can be very large and are not always easily recognized or removed, but can be reduced by establishing a relatively complete library of interferograms. Furthermore, the presence of vegetation, especially trees, can cause rapid decorrelation of images with the present SAR satellites. The 1-D range-change field determined through analysis of numerous interferograms, together with precise 3-D velocities from GPS networks and strain data (Bock and Williams, 1997), can provide a view of active deformation with unparalleled spatial and temporal resolution. Complete integration of the data will allow us to produce a continuously updated space-time image of slip and loading.

Currently we are dependent on a single European satellite mission to provide InSAR images of western North America. A consortium (WInSAR) has recently been established under the SCEC umbrella to coordinate the purchase and acquisition of InSAR data from representatives of the European Space Agency. It is clear that a dedicated U.S. InSAR mission would serve to alleviate many of the deficiencies described above (e.g., imaging at L-band frequencies to reduce decorrelation caused by vegetation) and to guarantee the availability and accessibility of these critical data. The PBO science goals would be advanced significantly if U.S. agencies were to go forward with a dedicated InSAR mission. In the absence of such a mission, the inclusion of InSAR data could be accomplished by the continued purchase of data, but this option would not satisfy all of the needs of PBO.

b. Airborne Imaging Platform (GeoSAR, ALSM)

(i) GeoSAR data can be acquired via a dual-frequency radar system mounted aboard a Gulfstream or equivalent jet, at a rate of approximately 1000 linear km per day. Airborne Laser Swath Mapping (ALSM) is acquired by mounting the imager aboard a helicopter or standard fixed-wing aircraft, acquired at about 200-300 km per day. Either system can produce the needed requirements: 10 km-aperture data with horizontal posting of about 5 m and vertical posting of about 0.3 m. Visible-spectrum tonal data in stereographic pairs would be acquired via a large-format color digital camera mounted aboard a low-stall-speed aircraft, at the same scale and along the same flight paths as the topographic data.

4. Seismometers

Crustal deformation is inherently linked to the occurrence of earthquakes along plate boundaries. In order to complement surface strain measurements, it is important to characterize strain at seismogenic depths. This cannot be directly observed by geodetic means, so we must resort to the use of strain indicators, i.e., phenomena related to strain, but which are uncalibrated. Presently, microearthquake activity offers the most straightforward way of doing this. The use of borehole seismometers provides the best means for detecting small events. These small events can be used to map detailed structure using relative relocations at tens of meter resolution.

Variations in seismicity or in the elastic properties of the medium can be interpreted as strain transients. For example, an analysis of the 11-year high-resolution microseismicity record at Parkfield revealed systematic spatial and temporal changes in the slip rate indicated by the repeat times of identical multiple events that were synchronous with other indicators of deformation. For the 2.5-year period beginning in October 1992, the analysis defined a deformation pulse of increased slip leading to and accompanying a series of M=4.6 to M=5 earthquakes that occurred during that time.

Accurate relocations of earthquakes and their interpretations within the PBO study area is an ongoing area of research that provides invaluable records of microseismicity.

Appendix B. Observing the Long-Term Deformation Field: Use of Geological Measurement

Surface geological constraints provide kinematic data with observational periods of less than a day (coseismic strain of historic events) up to millions of years. This perspective is critical to the success of PBO because the phenomena we wish to observe, including the seismic cycle of major plate boundary faults and intraplate fault zones and the dormancy periods of active volcanoes, are generally much longer than can be observed using geodetic instrumentation. With a systematic approach, the modern observational tools of active tectonic studies will create a map-view "movie" of the sequence of earthquakes and volcanic eruptions across the entire plate boundary zone, in time-slices ranging from hundreds to millions of years. The time series of earthquakes and eruptions on individual faults and volcanoes, and the measured deformation rates, will establish the geophysical context of the contemporary instrumental data. The interrelations between these events, such as correlation between earthquakes and volcanism, or migratory patterns of earthquakes or eruptive activity, will be observed in detail for the first time across any plate boundary system.

Here we illustrate through several case histories the measurement strategies and the quality of data that will be obtained. These case histories illustrate the need for three specific classes of information, including (1) remote sensing data, which establishes a template for (2) field-based displacement measurements, providing the spatial constraint (numerator) on kinematics, and (3) laboratory-based geochronologic measurements, providing the temporal constraint (denominator).

Strategies

Time series of earthquakes and volcanic eruptions are obtained from geological relations and geochronology in surface outcrop, trenched fault zones, and shallow boreholes. For example, where major fault zones interact with accumulating sediment, earlier earthquakes disrupt older layers, which are then overlain by undisturbed layers. The age of the earthquake is constrained by dating the youngest disturbed layers and the oldest undisturbed layers, most precisely with ^{14}C. A trench wall cut across the San Andreas Fault at Pallett Creek, California (Sieh, 1978), for example, documents the last ~ M8 earthquake in 1857, and a number of previous events ranging back to about 800 AD.

Using this technique on a number of trenches, it was possible to document the last ten events on this segment of the San Andreas. The age brackets of these events, which suggest approximately one surface-rupturing earthquake every 132 years, and a tendency for temporal clusters of two to three events, separated by dormancy periods of 200-300 years (Sieh et al., 1989). Selection of a locality for trenching, and measuring offsets of very young geological features, generally involves use of high-resolution topography and aerial photography, and in some cases shallow subsurface acoustic imaging, such as multi-channel seismic (MCS) reflection profiling or ground-penetrating radar (GPR).

Deformation rates, including slip rates of active faults and shortening rates across active fold belts, and eruptive fluxes of magma, are measured using aerial photography at about 1:10,000 scale and high-resolution topography, surface and shallow subsurface data (~1 to 100 m), and isotopic analyses. With the latest techniques, it is generally possible to measure slip rates on faults to a precision of 10-20% or better over periods ranging from as few as two to three earthquake cycles on up to hundreds or thousands. For example, along the San Andreas Fault near San Bernardino, the fault offsets a series of ancient river terraces of Cajon Creek by differing amounts. In one particular case an abandoned terrace riser has been offset by the fault, documented with 2 m-resolution topography. Erosion of the bank occurred 8,350 ± 950 years ago (from radiocarbon age of sediments deposited just after the riser formed), and it is offset 200 m, yielding a slip rate of 24 ± 5 mm/yr (Weldon and Sieh, 1985). Offsets of three additional geomorphic features of varying age, give similar results. Collectively these measurements show that slip on the San Andreas at Cajon Creek has been relatively steady over the last 15,000 years at 25 mm/yr, in reasonable agreement with geodetically determined rates across the same area.

Tectonic geomorphology provides a means to measure the often significant strain between major fault zones, or strain not accommodated by surface-breaking ruptures, particularly in areas of horizontal shortening adjacent to the San Andreas transform and Cascadia subduction zone. These methods make it possible to measure deformation rates and even earthquake time series. For example, dating of uplifted geomorphic surfaces along the Wheeler Ridge anticline in southern California (Medwedeff, 1992; Mueller and Talling, 1997) indicates rates of uplift, horizontal shortening, and lateral fault propagation over the past 200,000 years. In realms of active subsidence, such as the hanging walls of normal faults, deformed strata in the shallow subsurface record the strain across an area. This deformation can be imaged with GPR and shallow MCS reflection profiling. Calibration of the age of these strata with shallow boreholes provides the time correlation required to quantify deformation rates. In another example, along the Cascadia subduction zone, an earthquake in 1700 AD was discovered on the basis of coseismic submergence from northern California to southern British Columbia, changing coastal forest and marshlands into tidal mudflats, killing previously healthy vegetation (Nelson et al., 1995). The size of the affected area implies an earthquake in the range M8 to M9. Trenching studies along coastal drainages in the Pacific northwest suggest that at least three major events occurred along the Cascadia margin in the last 2,000 years (Atwater, 1992). Thus these methods provide information on both the time and size of earthquakes, even though it is not possible to observe the rupture plane directly.

Intraplate fault zones along the plate boundary are mostly short (ca. 50 km), normal and strike-slip fault segments and have slip rates and event frequencies an order of magnitude lower than major plate boundary transforms and subduction zones. Despite this difference, they account for some 20-30% of the total plate boundary strain budget and produce devastating earthquakes. Yet the relation of the intraplate system to the major plate boundary structures and volcanism, and the basic pattern of strain release within the system itself, are largely unknown. Obtaining precise earthquake time series is needed on these structures in order to observe the spatial and temporal pattern along and across the system. It has been possible to obtain the chronology of earthquakes from trenching along various segments of the Wasatch fault zone, Utah (Machette et al., 1992), which is considered one of the most active intraplate faults. Nonetheless, recurrence on these segments is on the order of thousands rather than hundreds of years.

One of the difficulties in studying the intraplate system is that the recurrence intervals for some faults and volcanoes may be as great as tens to even hundreds of thousands of years, out of the range of precise ^{14}C dating. Recent developments in surface exposure dating using short-lived cosmogenic nuclides such as ^{36}Cl, ^{20}Ne, ^{10}Be, and ^{26}Al have revolutionized tectonic geomorphology and volcanology, making it possible to date any deformed or offset geomorphic surface (such as a lava flow, alluvial fan or river terrace), and the fault planes themselves (bedrock scarps), without requiring carbonaceous material. These techniques will be of central importance for determining deformation rates and event chronologies for all plate boundary fault zones. For example, the intraplate Hebgen Lake Fault in the Yellowstone area, which last ruptured in an M7.5 earthquake in 1959, at one locality exhibits a 13-m-high bedrock scarp progressively exposed by multiple events. ^{36}Cl exposure-age dating up the scarp shows both the amount of slip and the times of earthquakes over the last 40,000 years (Zreda and Noller, 1998). The data suggest two clusters of events lasting a few thousand years, separated by hiatuses in strain release lasting 10,000 to 15,000 years.

These techniques have recently been combined with diffusion modeling of alluvial scarps, providing the capability to determine the event chronology and slip rate of any dip-slip fault cutting an alluvial scarp in the age range 10,000-500,000 years. Previously, scarp diffusion modeling was limited by highly variable diffusivity, or the ease with which the scarp erodes (Hanks, 1998). Surface exposure dating of vertical profiles of gravel samples taken from trenches across the faults provides the necessary temporal information to determine both diffusivity and the time since the last event or events (F.M. Phillips, written commun., 1999). Samples are taken from a "control" position on the alluvial surface, and from the erosive and depositional regimes on either side of the scarp. The diffusion equation for the scarp and the production equations for ^{36}Cl are then combined, yielding model profiles. For the intraplate Socorro Canyon Fault, New Mexico, the control profile indicates that the alluvial surface formed 140,000 years ago. The scarp ages and model indicate the fault has slipped at a rate of 0.03 mm/yr since then, with one rupture 130,000 years ago and a second 40,000 years ago, with scarp diffusivity 0.4 m^2 kyr^{-1}.

Using Pre-Quaternary geologic features such as Late Cenozoic basins and large-scale geologic markers such as isopachs and thrust faults, it is possible along much of the plate boundary to reconstruct the strain history on the 1-10 Myr, with increments of strain resolved in 1-4 Myr intervals. This approach has been applied to the central part of the Basin and Range province (**Fig. 4**), and yields the Late Cenozoic motion of the Sierra-Great Valley block, a large, relatively undeformed crustal fragment along the plate boundary (Wernicke and Snow, 1998). This type of reconstruction may be extended to include the entire plate boundary fault zone, permitting finite vector displacement fields to be determined for the Late Cenozoic. The vector fields so derived may be compared with the contemporary field determined from GPS and earthquake time series, and with constraints on overall plate motion derived from plate tectonic reconstructions (e.g., Atwater and Stock, 1998). Like the field derived from GPS, this field is a prime constraint on any physical model of plate boundary deformation (e.g., Shen-Tu et al., 1998).

Recommendations

(1) On-line, high-resolution topographic and tonal data along active plate boundary deformation zones and eruptive centers. Approximately 25,000 km of 10 km-aperture data would be required

to produce a systematic topographic database at roughly 1:10,000, with horizontal posting of about 5 m and vertical posting of about 0.3 m. Coverage would generally not include areas without documented Quaternary deformation, and will cover roughly 5% of surface area of the plate boundary deformation zone. Ordinarily, obtaining topographic data of the quality along Cajon Creek and high-resolution tonal data (either digital or analog photography) are the rate-limiting factors in making these measurements. In the case of the Cajon Creek study, a vigilant search was required to obtain adequate aerial photography (generally in the 1:10,000 range), a search that frequently ends in failure, because these data are usually of insufficient quality, and are haphazardly acquired and archived by numerous private contractors and government agencies. The topography was painstakingly acquired in the field using plane table methods. A top priority for this element of PBO is therefore a systematically acquired and archived community database. These data will include either radar-based (e.g., GeoSAR) or laser-based (e.g., Airborne Laser Swath Mapping, ALSM) topographic imaging with vertical posting in the 10-30 cm range, and high-resolution digital imaging in the visible range at low sun-angle with pixel size in the 10-20 cm range. These data, once available, will remove a major existing barrier to obtaining long-term kinematic measurements, because acquiring them for any single project is prohibitively expensive and time-consuming. In addition to the economies of scale realized by systematically imaging all major plate boundary structures, this unprecedented database will likely catalyze interest in PBO within the broader Earth science community.

Appendix C. The Backbone

The PBO backbone array, composed of continuous GPS sites, serves several functions. First, it will sample the long-wavelength decadal velocity field of deforming western North America. In several areas, including significant parts of Alaska, western Canada, Mexico, and the interior northwest, the backbone array measurements will be the first systematic high-precision geodetic measurements to be made, and there is a strong possibility of serendipitous discovery. In regions where there are no measurement clusters, the backbone will provide the only PBO measurements for that region. Second, even in better-characterized areas the backbone array will provide the regional framework on which to hang more detailed investigations. Deployment of a robust backbone array will allow the analysis of the GPS data to be separated into regional sub-arrays that can be combined together using common backbone sites with essentially no loss of precision. Third, the backbone will provide a common reference of permanent sites to link together dense survey-mode measurements, which will continue to be made in many areas. These last two functions can be combined as the 'infrastructure' function of the backbone array.

To fulfill the functions listed above, the backbone array needs a station spacing on the order of 100-200 km, with each function having a slightly different requirement. In terms of sampling the velocity field, the denser the measurements the better. This is most true in the areas where less is known about the deformation field at present, and a reasonable strategy would be to make the backbone denser in these areas. Even in well-sampled regions, however, the backbone array cannot be neglected. For example, over about one third to one half of California the velocity field is already well known through existing permanent sites at a 100-km spacing, but in the remainder of the state (including the southern Cascadia margin) the details of the velocity field remain unsampled. The infrastructure requirements of the backbone array are less strict. Two practical criteria can be defined: (1) the backbone should be dense enough so that even if the closest site is unavailable, reliable ambiguity resolution can be carried out for any densification measurement made within the array; (2) for the purposes of combining regional solutions into a single PBO-wide solution, it is necessary to include several backbone array sites in every regional solution, and desirable that such regional solutions generally span less than several hundred kilometers by several hundred kilometers to minimize errors caused by subtle differences in software, reference frame definition, and other inconsistencies between analysis centers. Both of these criteria suggest that about 300-km spacing would be an upper limit for the backbone array and < 200-km spacing would be ideal. The ideal spacing will be modified by logistical realities in a number of areas. For example, over large areas of Alaska and western Canada, 100-km spacing is simply not feasible, and it will be difficult to achieve at reasonable cost in other remote areas.

Deployment of the backbone array will be a challenge mainly because of the vast area to be covered by the PBO. As a result, it is important to begin the process early, ideally even before the PBO is formally approved. Identification of potential backbone array sites should begin as soon as the final design spacing for the backbone has been decided. This initial effort will be most efficiently done in many regions by local groups, either scientists already involved in running continuous GPS arrays or experienced P.I.s not presently working on permanent arrays. Once initial site selection has been completed, the detailed site selection and permitting can be done by the appropriate installation contractor(s) (see Management, Section 4.0). It is critical to begin this process early, and to make use of local expertise, since the process of obtaining permits can be quite lengthy and will limit the ability to efficiently install the backbone array. For example, it is almost always much more efficient to send an installation crew to a given region once to install 10 sites than it is to send the crew there several times to install one or two sites each time. This is an area where a modest investment on the part of the NSF prior to approval of the main PBO funding will pay off handsomely in increased installation efficiency when PBO is deployed.

Appendix D. Deployment in Canada and Mexico: International Collaboration

The North American plate boundary zone under consideration passes through Canada and Mexico, as well as the U.S. The problems that we seek to study show no such boundaries. The San Andreas Fault system extends into Mexico. The cordilleran structures seen both in Alaska and in the Pacific Northwest of the U.S. extend without interruption into Canada. Thus it is of great importance that the U.S. closely collaborates with its Canadian and Mexican counterparts for a fully integrated Plate Boundary Observatory.

Mexico

Mexico is characterized by a wide variety of contemporary tectonic processes including ocean spreading in the Gulf of California, continental transform faulting along the Cerro Prieto-Imperial Faults, extension in the Basin and Range of the Sonora desert, subduction along the Middle American trench, and extensive volcanic/geothermal activity along the Mexican volcanic belt and Colima rift. A thorough understanding of Pacific/North American plate boundary deformation, in particular the transition from ocean spreading in the Gulf of California to continental transform faulting along the southern San Andreas Fault, requires coordinated observations in Mexico and close collaboration with our Mexican counterparts. Understanding earthquake hazards in southern California also require observations in northern Mexico and Baja, California. Faults extend directly across the international border such as the Imperial-Cerro Prieto Fault system, the Elsinore-Laguna Salada Fault, the San Miguel-Vallecitos Fault which likely transfers strain to the Rose Canyon Fault in San Diego, and the trans-Baja Faults such as the Agua Blanca which transfers some portion of the plate motion to southern California offshore faults (e.g. San Clemente). Furthermore, Press and Allen (1995) suggest that small changes in spreading in the Gulf directly influence earthquake activity in southern California.

Mexican and U.S. investigators have been collaborating on survey-mode GPS projects in Mexico since the 1980's, including deformation studies in southern California/Baja/Gulf of California, regional deformation along the Pacific/North American plate boundary, deformation of the Jalisco block associated with subduction of the Rivera and Cocos plates, and volcanic deformation. Several large and destructive earthquakes in Mexico during this period have stressed the societal importance of these tectonic studies. Mexican participation comes primarily from two groups, Centro de Investigacion Cientifica y de Educacion Superior de Ensenada (CICESE) in Ensenada, Baja, California, and Instituto de Geofisica, Universidad Nacional Autonoma de Mexico (UNAM) in Mexico City. These two groups are also responsible for operating extensive arrays of seismic instrumentation.

Permanent geodetic (continuous GPS) installations in Mexico have been scarce, limited to sites in Baja, California, in Ensenada, a volcano near Mexico City, and a site in the Jalisco block. They have, however, received a large boost in the last 1-2 years, with significant funding from CONACYT (the Mexican National Science Foundation) to UNAM and CICESE, sufficient for the establishment of about 15 continuous GPS sites (see: http://tlacaelel.igeofcu.unam.mx/

~vladimir/gpsred/gpsred.html), as well as receiver contributions and technical support from the Southern California Integrated GPS Network (SCIGN). SCIGN has contributed GPS equipment to CICESE (2 units) and UNAM (1 unit) and assisted in the installation of sites with CICESE in northern Baja, California (San Pedro Martir), and with UNAM at Yautapec in central Mexico. The UNAM group has recently installed a new site in Acapulco (using the installation at Yautapec as a model) as part of its collocation of continuous GPS at broadband seismic stations. The CICESE group has performed site reconnaissance at two islands off of Baja, California, Isla Coronado and Isla Guadalupe (the latter an important site firmly situated on the Pacific plate), and at Cucapah, Puerto Peñasco, and Hermosillo. Sites established with CICESE will be fully integrated within SCIGN. In addition, a 14-station national GPS network (Red Geodesica Nacional Activa - RGNA) is operated by Instituto Nacional de Estadística, Geografía e Informática (INEGI) with older generation GPS equipment. Scripps Institution of Oceanography is working with INEGI to upgrade their equipment, and has contributed a new receiver for operation at INEGI's central facility in Aguascalientes. A continuous GPS training session at SIO was held for participants from the 3 primary Mexican groups (CICESE, INEGI, and UNAM) on January 11-14, 2000, covering site monumentation and installation, receiver operations, archiving, data dissemination, and data analysis.

We thus expect strong involvement in PBO by these Mexican institutions and other scientists interested in plate boundary deformation.

Canada

Western Canada, stretching from the foothills of the Canadian Cordillera in the east to the active volcanoes of the Cascades and the offshore ocean-spreading centers along Pacific coast in the west, from the Cascadia Subduction Zone in the south, to Canada's highest mountains in the St. Elias region of the southwest Yukon Territory in the north, is the most earthquake-prone region of Canada. The present-day western North American plate boundary extends for over 1000 km along Canada's west coast. Southwestern coastal British Columbia marks the northern terminus of the Juan de Fuca/North American convergent zone and, as such, the continuation of the Cascadia megathrust rupture zone which last slipped in a magnitude 9 event in 1700. Central Vancouver Island has also experienced two large (M>7) crustal earthquakes in the past century. The relationship of these large shallow events to the subduction thrust process, their recurrence interval, and their potential geographic distribution are not well known and yet, such earthquakes comprise the major seismic hazard in the Pacific Northwest. The Tuzo Wilson Knolls located northwest of Vancouver Island mark the triple-junction where convergent plate motion changes to predominantly transform motion between the Pacific and North American plates along the Queen Charlotte-Fairweather Fault system. Canada's largest historic earthquake (magnitude 8.1) occurred on August 22, 1949, rupturing a 500 km segment of the northern portion of the Queen Charlotte Fault (QCF). Further north, the St. Elias region, straddling northwest British Columbia, the southwest Yukon Territory, and southeast Alaska is characterized by intense seismic activity and uplift rates exceeding 3 cm/yr caused by the transition of transform motion to convergence and subduction along the Aleutian Trench. This area of the plate margin has experienced many large earthquakes in the recent past, including a sequence of three earthquakes of magnitude 7.4 to 8.0 in the year 1899, and a magnitude 7.9 earthquake which occurred along the Fairweather Fault in 1959. Earthquake activity in Canada's northern regions is not confined to the coastal margin but also extends inland. A significant zone of seismicity follows the Dalton and Duke

River segments of the Denali fault zone through the southwest Yukon, there is minor seismicity between the Denali and Tintina Fault system to the northeast, and the rate of seismic activity increases again at the eastern edge of the northern Cordillera within the Mackenzie Mountains of the Northwest Territories and the Richardson Mountains of the Yukon Territories.

Canada has an active program of seismic and deformation monitoring that provides fundamental infrastructure which can be expanded and densified in support of the PBO. This program is carried out by the Geological Survey of Canada (GSC) at the offices of the Pacific Geoscience Centre (PGC) located in Victoria, British Columbia. With the close cooperation of the USGS, measurements of horizontal crustal strain were initiated on Vancouver Island in 1981 using laser-ranging techniques which were later replaced by cooperative GPS field campaigns spanning the international border between northwest Washington and southwest British Columbia. The establishment of a network of continuous GPS stations called the Western Canada Deformation Array (WCDA) (see http://www.pgc.nrcan.gc.ca/geodyn/wcda.htm) was initiated by the GSC in 1991. The WCDA currently consists of 9 sites located in southwestern BC at a nominal spacing of about 150 km. Within this broad spacing, GPS field campaigns are periodically repeated to map details of the strain field. The WCDA served as the impetus for the establishment of the PANGA CGPS array in Washington and Oregon and together they now comprise 30 sites for the study of crustal motions along the Cascadia margin. The GSC also operates a continuous GPS site at Whitehorse whose data have been used extensively for crustal deformation studies in SE Alaska undertaken by the University of Alaska's Geophysical Institute.

Canadian scientists have participated in the PBO workshop and have expressed strong interest in becoming international partners in PBO science and technology. Canadian researchers are optimistic that success of PBO in the U.S. can be leveraged to expand monitoring in Canada, especially in the densely populated urban areas along the Cascadia subduction boundary and the extremely active transform boundary along the outer coast of northwest British Columbia.

Appendix E. Education and Outreach

Background

Because plate boundary deformation zones can be broad, extending thousands of kilometers into continental interiors and accounting for 15% of the Earth's surface, they should be a critical area of study, both from scientific and societal points of view. A Plate Boundary Observatory (PBO) established along the Pacific/North American plate boundary would measure deformation over a broad spectrum of spatial and temporal scales and offer an opportunity to combine several tools and techniques based on existing and new technologies (GPS, strain, seismic, and INSAR measurements). The data collected will greatly improve knowledge of such transients as post-seismic deformation, increased strain due to large earthquakes, slow earthquakes, and multiyear aseismic slip.

The PBO is being proposed as a coordinated and integrated project with the proposed USArray and the San Andreas Fault Observatory at Depth (SAFOD) initiatives as part of the EarthScope initiative.

Several fundamental questions that will be addressed by PBO scientists will be of interest to the public, e.g.,

- How does plate motion ultimately produce an earthquake?
- How do faults interact?
- How do earthquakes interact?
- What fraction of fault slip is aseismic?
- What kinds of earthquake-related transients are there?
- Do pre-event transients exist that may be utilized for forecasting?
- How do volcanic eruptions occur?
- Can volcanic eruptions be predicted?

Rationale for Public Awareness and Informal Education Programs associated with PBO

We suggest a Public Awareness and Informal Education program for PBO that complements the Education and Outreach (E&O) efforts of the proposed USArray and the San Andreas Fault Observatory at Depth (SAFOD) initiatives. This approach will minimize duplication and encourage coordination among all groups.

- The western United States (PBO region) has several large urban populations with exposure to volcanoes and earthquake-related risks (e.g., Los Angeles, San Diego, San Francisco, Portland, Seattle, Salt Lake City). PBO presents the opportunity to reach these populations (totaling over 25 million people) through museum displays and media (TV, radio, videos, web and print) coverage. An informed public is critical to many societal and environmental decisions related to earthquake hazard preparedness, mitigation, and response, such as land use and building code issues.

- Public awareness programs would promote broader interest in and support for science, especially Earth science.
- Public awareness programs would improve the image of Earth science and scientists.
- Public awareness programs are an extremely effective mechanism for adult education.

Outreach Products Proposed for PBO

The PBO public program should be designed to complement the outreach efforts of the USArray and the San Andreas Fault Observatory at Depth (SAFOD). Primary emphasis will be on informal education and public awareness programs. An advisory group consisting of PBO scientists and end users will provide guidance in development and construction phases of the products, as well as ongoing assessment and feedback.

Museum exhibits and programs

A single exhibit, developed, for example, with California Science Center and/or Exploratorium education staff, could be duplicated at other museums or could be introduced as a traveling display throughout the life of the project. The exhibit would utilize new techniques employed by progressive museums, such as hands-on learning tools or components, web-casting, on-site videotaped studio shows, or live interviews, etc.

Informal education / public awareness products

- PBO press kits: General fact sheets, contact information, special announcements, related informational materials. Regional press kits: Each kit will be designed to serve information needs of specific regions along the plate boundary.
- Video(s): 5-15 minute-length beta videos that can be used by museums, media reporters, civic groups, and schools. Research marketing opportunities / incorporation into future educational TV projects such as Discovery Channel's series on natural disasters / phenomena.
- Posters: hard copies for widespread distribution; electronic versions for posting to a PBO web site.
- Distribution of public education posters / materials at appropriate sites along public highways located near the plate boundary (I-5, I-15, etc.).
- Radio: "The PBO Minute" – investigate potential for radio infomerical – could tie with education materials located along highway sites like the successful "LA Underground" aired by KFWB in Los Angeles.

Web site

We suggest capitalizing on work already accomplished by several groups with existing web sites, by creating pages that can be integrated into existing sites, or by creating a new web site that is linked to all existing, related web sites. Example web sites would be USArray, SAFOD, SCEC, SCIGN, IRIS, etc.

We suggest creating add-on components to existing web-based SCEC and SCIGN education modules.

Structure and content of a web site will be determined by Outreach staff and PBO scientists, with advice from a select group of end users in the community-at-large. Possible items to include:

- General info on PBO – and related projects such as USArray and SAFOD
- Time history of deployment, papers, high profile results, science questions
- Site information - photos, words, technical info
- Acknowledgements and contacts
- Links to data, data return statistics
- Maps – in appropriate time increments
- News pages
- Links to educational materials and resources
- Calendar of related workshops, seminars

Access to Data

Web tools and software for viewing / downloading data.

Interactions with Other E&O Efforts

These would include collaborations with the USGS, state surveys, national parks and forests, as well as the extension of SCIGN educational modules (middle school, high school, college).

Appendix F. White Paper Contributors

Besides the PBO Steering Committee, several individuals made significant contributions to the White Paper.

- Extended discussion papers on PBO topics:

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- Presentations at the workshop:

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