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The region of Axial Volcano, Juan de Fuca Ridge region provides an excellent opportunity to study the interplay between active "hot spot" and "mid-ocean ridge" magmatic systems. Important questions include how the two magma systems are fed; their magma and heat budgets; the degree of interconnectedness (or interaction) between them; their relationship to seismicity and geodetic strains; the role of each in plate-tectonic spreading and and crustal formation; and their effect on the geochemistry (e.g. mixing, fractionation) of erupted basalts. Information on the physical layout of the magma systems is critical to the study of each of these issues. The purpose of this research was to investigate these questions through the tomographic imaging of the region using seismic data from an active seismic airgun-to-obs experiment. The experiment was remarkably successful, both in the sense that voluminous high-quality data were obtained, and in the sense that very clear signials associated with magma were detected in that data. The key elements of the new three-dimensional compressional velocity model of the Axial and Coaxial magma systems are (West 2001; West et al 2001; Menke et al. 2002):

1. **A Very Large Axial Magma Chamber.** At a depth of 2.25 to 3.5 km beneath Axial caldera lies an 8 by 12 km magma chamber containing 5-25% melt (West et al. 2001). At depths of 4-5 km beneath the sea floor there is evidence of additional melt, in lower concentrations (a few percent) but spread over a larger area (Fig 5). Residence times of a few hundred to a few thousand years are implied (West et al. 2001).
2. **A smaller Coaxial Magma Chamber, unconnected with the one at Axial.** The magma chamber is loacted at the "Source Site" of the 1993 eruption (Menke et al., 2002). It is at least 6 cubic km in volume and contains at least 0.6 cubic km of melt, enough to supply at least several eruptions of size equal to the one in 1993.
3. **Several other small low velocity zones are possibly outlier magma chambers from Axial.** Two other low-velocity zones occur in the shallow crust near Axial volcano, one about 10 km north of the caldera on the North Rift, and the other about 10 km south of the caldera but displaced to the west of the South Rift (West 2001). They appear unconnected to the main Axial magma chamber and might possibly represent small accumulations of melt left over from past lateral diking events.
4. **Strong thickening of the crust beneath Axial volcano.** The crust thickens from about 6 km far from Axial to 8 km near Axial to 11 km beneath the summit (West 2001).

Publications:

1. Menke-W, Shallow crustal magma chamber beneath the axial high of the Coaxial Segment of Juan de Fuca Ridge at the "Source Site" of the 1993 eruption, in press in Geology, 2001.
2. West-M, The deep structure of Axial Volcano, Ph.D. Thesis, Columbia University, 2001.
3. West, M., W. Menke. M. Tolstoy, S. Webb and R. Sohn, Magma storage beneath Axial volcano on the Juan de Fuca mid-ocean ridge, Nature 25, 833-837, 2001.

Data and other products This project collected new data, which is freely available on-line at http://www.ldeo.columbia.edu/user/menke/AX/. Some software, including the tomograhy code, that was written for the project is available at http://www.ldeo.columbia.edu/user/menke/software/.

Introduction.

Axial volcano, in the Northeast Pacific, is a large ridge-centered seamount associated with the Cobb-Eickelberg hot spot. Its position on the actively-spreading Juan de Fuca ridge (JdF, 60 mm/yr full spreading rate), its proximity to western North America, its shallow (1600 m) summit depth, its prominent 3x8 km wide caldera (figure 1A), its high heat flow and its vigorous hydrothermal activity have led to its being the focus of numerous research efforts (e.g. the special sections in the Journal of Geophysical Research (vol. 95, 1990), Geophysical Research Letters (vol. 22, 1995 and vol. 26, 1999). Two recent volcanic eruptions in the area (described below) attest to the vigorous magmatic activity of Axial and the nearby Coaxial segment of the Juan de Fuca ridge.

In 1993 a large sea floor volcanic eruption occurred along the Coaxial segment of the JdF, immediately to the northeast of Axial Volcano. This eruption was detected during its early stages by hydroacoustic observations (Dziak et al., 1995), and subsequently studied intensively (e.g. special section in January 15, 1995 issue of Geophys. Res. Lett.). The eruption appears to have been caused by the lateral propagation of a dike from the magma chamber of on the southern part of Coaxial segment to a site 25 km to the northeast (Dziak et al., 1995). The sequence of events seems to be similar to the 1974-1985 rifting episode in northern Iceland, which involved the lateral propagation of dikes away from Krafla Volcano (Brandsdottir and Einarsson, 1979). The Iceland rifting episode led to about 9 meters of spreading of the North American - Eurasian plate boundary. The amount of spreading associated with the Coaxial eruption is not known.

In 1998 a second eruption occurred in which a dike propagated from near the Axial caldera to a point about 50 km to the south (Dziak and Fox 1999). The propagation of this dike was also monitored by hydroacoustic means (Dziak and Fox 1999). This eruption caused 3 m of subsidence of the Axial caldera floor (Fox 1999). Geological mapping of lava flows along Coaxial and their chemistry, which is distinct from Axial basalts, have been used to argue that the sources of the Axial and Coaxial lavas are distinct (Embley et al. 2000).

This region thus provides an excellent opportunity to study the interplay between active "hot spot" and "mid-ocean ridge" magmatic systems. Important questions include how the two magma systems are fed; their magma and heat budgets; the degree of interconnectedness (or interaction) between them; their relationship to seismicity and geodetic strains; their role of each in plate-tectonic spreading and and crustal formation; and their effect on the geochemistry (e.g. mixing, fractionation) of erupted basalts. Information on the physical layout of the magma systems is critical to the study of each of these issues. Such a model, based on tomographic imaging using seismic data from an active seismic airgun-to-OBS experiment that we performed in 1999, is now available. The key elements of this three-dimensional compressional velocity model of the Axial and Coaxial magma systems are (West, 2001; West et al. 2001, Menke et al. 2002):

1. **A Very Large Axial Magma Chamber (figure 1B).** At a depth of 2.25 to 3.5 km beneath Axial caldera lies an 8 by 12 km region of very low seismic velocities (figure 1C,D) that can only be explained by the presence of magma (West 2001; West et al. 2001). In the center of this magma chamber the crust is at least 10-20% melt. At depths of 4-5 km beneath the sea floor there is evidence of additional melt, in lower concentrations (a few percent) but spread over a larger area. The total volume of the magma chamber is about 200 cubic km, of which 5-26 cubic km is melt. This large volume of magma, compared with that erupted in 1998, imply residence times of a few hundred to a few thousand years (West et al. 2001).

2. **A smaller Coaxial Magma Chamber, unconnected with the one at Axial.** The magma chamber is loacted at the "Source Site" of the 1993 eruption (Menke et al., 2001). It is at least 6 cubic km in volume and contains at least 0.6 cubic km of melt, enough to supply at least several eruptions of size equal to the one in 1993. No mid-crustal connection of this magma chamber with the magma chamber of nearby Axial volcano is evident, confirming previous geochemical and geological

studies that argued against mixing between the two. The lack of connectivity implies that magma transport though the uppermost mantle and lower crust are very highly focused into narrow (<5-10 km) conduits.

3. **Several other small low velocity zones are possibly outlier magma chambers from Axial.** Two other low-velocity zones occur in the shallow crust near Axial volcano, one about 10 km north of the caldera on the North Rift, and the other about 10 km south of the caldera but displaced to the west of the South Rift (West 2001) (figure 2B). They appear unconnected to the main Axial magma chamber and might possibly represent small accumulations of melt left over from past lateral diking events.

4. **Strong thickening of the crust beneath Axial volcano.** The crust thickens from about 6 km far from Axial to 8 km near Axial to 11 km beneath the summit (West 2001). The long-wavelength 6-8 km thickening is consistent with predictions based on gravity data (Hooft & Detrick 1995). The shorter wavelength 8-11 km thickening, which creates a three km thick crustal root beneath the volcano, is not predicted to have an observable gravity signature. A sharp, normal Moho boundary is detected at the base of the crust (including at the base of the root).

Proposed Research

At present, the integration this new understanding of the magmatic structure of Axial volcano with other geophysical data has been largely qualitative, which is ironic given that they provide a very quantitative and detailed description of the subsurface. I therefore propose to develop a hierarchy of quantitative models of the region that make quantitative predictions about the state of stress that can be compared with seismicity and geodetic data. The release of pressure within a magma chamber during an eruption causes changes in the state of stress within the surrounding rock (and hence possibility to a change in seismicity), geodetic displacments of the ocean floor (e.g. subsidence and tilt), and changes in the pressure in neighboring, unconnected magmatic systems. These models will both predict these effects and assess how the uncertainty in the underlying tomograhic model effects these predictions. I expect that the very large lateral gradients in material properties will have a major effect on the stress field, and will give rise to phenomenon that would not be accurately modeled with the simple homogeneous halfspace models that are sometimes used in connection with volcanos.

A three-dimensional model of a volcanic system, especially one based on a tomographic inversion containing thousands of parameters, could in principle be very complicated. We can naturally ask, first which features in the model are critical for predicting a given phenomenon, and second whether these prediction are sensitive to errors that arise because of noise or poor resolution in the tomography.

We will address the first issue – the effect of model complexity - by developing a hierarchy of models that incorporate have different degrees of complexity.

Phase 0. One-dimensional models based on a homogeneous halfspace approximation with very simplified point and planar pressure sources representing the magma chamber and ridge segments, respectively (Mogi 1958). These are the "control" models to which more complicated models can be compared. Several features of the Axial tomographic model suggest that this control model will prove inadequate, especially in explaining "near field" observations. In particular, the shape of the magma chamber is quite complicated. It has a mid crustal extension to the southwest (Fig 2C), a wide deep-crustal base (Fig 2D, Fig 5) and a shallow ring dike structure (Fig. 1B).

Phase 1. A quasi-static 3D elastic model using the cavity assumption. Here the crust is modeled as a heterogenous elastic solid containing irregularly-shaped compressible fluid-filled cavities whose shape is inferred from the tomographic model. Both magma chambers and dikes can be modeled with such cavities. Plate-tectonic extensional stress can also be imposed over the whole region as a boundary condition. The key limitation of this model is that it does not allow stress-relaxation due to the finite viscosity of both the hot crustal material and of the magma itself.

Consequentally, this model only can be applicable to time scales that are short compared to the relaxation time of the crust as a whole (decades to centuries) but long compared to the viscous relaxation time of the magma (minutes to hours). It will thus be useful for examining processes that occur, say, in the days to years following and eruption, a time period for which many observations are available (discussed further below).

We will use the inexpensive, commercially-available *Beasy* analysis code (see http://www.beasy.com) for these Phase 1 stress calculations. Beasy is based on a boundary-element method. The earth is divided into many small regions delimited by surfaces composed of triangular (or quadilateral) faces. Each region must be homogenous, but adjacent regions can have arbitrarily properties. These surfaces can be shaped to match sea floor bathymetry, the depths to elastic moduli interfaces (infered from isovelocity surfaces with the tomograhic model) and the surface of the magma chambers. Stress and displcaement can be calculated both on the surface itself and at selected points in the interior of the regions. This code has proven reliable for a variety of stress analyses in a geophysical context (e.g. ten Brink et al. 1996 (see our figure 3), Gudmundsson et al. 1997). Its status as a “well-tested” code will also be important in the context of the error analysis discussed below.

We have some experience with Beasy at LDEO dating from the mid and late 1990's, when several of our students (e.g. D. Bohensteil) used it for crustal stress modeling. However, we will be requesting funds for renewing our software license, which expired in 2001.

Phase 2: Viscoelastic modeling. Magmatic events, such as dike emplacement, cause an immediate elastic response in the stress field followed by a slower change as viscous relaxation takes place in regions (such as the lower crust) that are hot enough to creep. This effect has been observed in Iceland, where creep is still going on decades after the Krafla rifting episode of 1976-84. Modeling of the Iceland measurements have been used to determine the viscous structure of the crust there (Pollitz and Sacks' 1996, Foulger at al. 1996).

We will use Jishu Deng's FEVER (Fine Element code for Visco-Elastic Rheology) software for the viscoelastic modeling. This code has been used successfully to model stresses in southern California (e.g. Deng et al., GRL 26, 3209-3212, 1999). Dr. Deng has graciously provided us full access to the code, and has agreed to informally collaborate with us in its use. The code uses a Maxwell solid representation of rheology, and solves for stresses on an irregular finite-element grid.

The reason for employing both elastic and viscoelastic models is that while the later better captures the overall physics of the deformation process, it is also requires knowledge of viscosity, a rather poorly determined property. Comparisons between the models will help us to identify conditions which require the more complicated analysis. Furthermore, some of the error sensitivity analysis that we will perform require numerous runs of the modeling code. As quasi-static calculations are much faster than viscoelastic ones, there is a practical advantage of being able to make use of them whenever possible.

Analysis of Error. Our goal is to be able to make quantitative geophysical predictions and to have some way of assessing their uncertainty. Three type of error are important: 1. Errors introduced from **scaling** from compressional velocity – the parameter derived from the tomographic model – and the elastic moduli and viscosity used in the modeling; 2. Errors associated with the reparameterization of the three dimesional model from the linear tetrahedral splines representation used in the tomography to different parameterizations used by the stress-analysis code 3. Errors associated with errors in the tomography, which include short-wavelength **noise** resulting from over-fitting the underlying traveltime data and from the limited **resolution** of the experiment. We briefly describe the role of each:

Scaling errors. In order to scale compressional velocity into elastic moduli, one needs estimates of the density and Poisson's ratio (or alternatively, the shear velocity). Data for the subsolidus part of the crust, which constitutes by far the largest part of the Axial model, is pretty good. Density scales well with compressional velocity. Hooft and Detrick (1993), for instance, successfully employ a simple linear relationship. Empirical studies in Iceland – which is similar in genesis to Axial volcano - indicate considerable uniformity of Poisson ratio, with a value of about 0.27 (Staples et al. 1997), even quite deep in the crust and close to volcanic systems. Data for fully liquid basalt melts are also available (e.g. see Decker 's 1975 bibliography). Their density is well-determined and, being liquids, they support no shear stresses. Physical properties (especially the Poisson ratio) at near-solidus temperatures, where shear wave velocities begin to fall, are less precisely known. Here the effect of temperature and melt fraction are very important. While the variation of Poisson's ration (or shear velocity) with temperature and melt study has received considerable attention (e.g. Sato et al. 1988, Faul et al. 1994), its functional relationship still has considerable uncertainty. Hence we believe that the greatest error in scaling are likely to be associated with the modeling of the near-molten and partially-molten material near and within the magma chambers.

Estimates of viscosity are necessary for the viscoelastic modeling. Numerous studies of the viscosity of magma are available (e.g. Wright et al. 1976). Our knowlegde of the viscosity structure of subsolidus basaltic crust comes from both geodetic observations (e.g. Pollitz and Sacks' 1996. work in Iceland) and through laboratory measurements of rheologic parameters. Kirby's (1987) nonlinear stress-strain law has formed the basis for numerous numerical models of crustal deformation, including at mid-ocean ridges (e.g. Chen & Morgan, 1990, Chen and Phipps Morgan, 1996). One extracts an effective viscosity from it by stipulating a strain rate and a temperature. Of the two, temperature is by far the most problematical, since there are no direct measurements of the thermal structure of the crust at Axial. Some information about the temperature structure can be gained through thermal modeling (e.g. Henstock et al. 1991, Phipps Morgan and Chen 1993, Cochran and Buck, 2001). West et al. (2001) assemble a thermal model for Axial using the surface of the magma chamber as a proxy for solidus isothermal surface with the temperure elsewhere constrained by Henstock et al.'s (1991) model. This method provides a reasonable extimate of the spatical variation of temperature, but one with considerable uncertainties.

Parameterization errors. The tomographic imaging was performed using Menke's (2002) raytrace3d code. The compressional velocity is represented on a tetrahedral grid, and a linear function is used to interpolate within each tetrahedron. Although tetrahedral grids can in general be quite "disorganized", this one was not. The nodes were arranged on a series of fairly smoothly-curved surfaces that paralleled the overall bathymetry of the region. The surfaces are not isovelocity, but far from the magma chambers they are nearly so. Successive depth surfaces had nodes at the same horizontal positions, so that they also formed six-faced polyhedra. This regularity simplifies the regridding process. The key issue is not the arrangement of nodes, but rather the type of interpolation used between. Both the Beasy and FEVER codes can handle the tomography's pattern of nodes, but they used different interpolation schemes. Beasy requires each polyhedron to be homogeneous. FEVER's is a polynomial representation. Thus some sort of regridding is necessary to make the different representations as similar as possible and in particular to preserve spatial gradients in physical parameters.

The Phase 1 elastic modelng requires an additional choice associated with the cavity model. What is in reality a continously-varying rheology with stiff/cold and soft/hot regions must be recast into distinct elastic and invicid fluid parts.

Tomography Errors. All attempts at geophysical inversion, including the Axial tomography, suffer from the well-known trading off of variance and resolution. All attempts to improve resolution, so as to be able to detect small features within the earth, inevitably increase the variance of the results. In seismic tomography, this usually translates into small-wavelength “chatter” in the velocity images that is related to “overfitting” the traveltime data. Similarly, all attempts to reduce the level of noise inevitably decrease the resolution of the image. Furthermore, the resolution is typically quite variable spatially, being poor in parts of the model (e.g. the bottom) that are undersampled by seismic rays.

However, not all velocity errors will effect the elastic modeling equally. Mid crustal velicity variation of a few percent – often the object of interest to tomographers – will not cause more than an insignificant few percent error of error in the calculated stresses. It is the tomography's ability to define the major elastic and viscoelastic features that is at issue. These include the bottom of the upper crust, the surface of the magma chambers and the depth at which viscosity rapidly decreases. Our feeling is that the long-wavelength (<5 km) variations in thickness of the upper crust is well-determined by the tomography, at least in the central region of the model, near the magma chambers. Some of the shorter wavelength variation is probably noise. We will need to assess its effect on the surface stress field. Our feeling is also that the location of the edges of the mid-crustal part of the magma chambers are well determined, since they are defined by sharp jumps in the traveltime curve. The thickness of the transition zone from fully solid to fully molten, however, is probably poorly determined. West et al. (2001) derived an estimate of Axial melt fraction (Fig 2D) using a combination of thermal models and tomography, following a methodology similar to the one used by Dunn et al. (2000) and Canales et al. (2000). They find relatively low melt fractions (< 5%) in the outer part of the magma chamber, which imply that these regions may be able to support shear stresses.

Errors in the tomography of the velocity structure of the interior of the magma chambers and errors in the scaling between velocity and elastic and viscous properties are conceptually distinct. But in practice they are interchangeable, since they occur in the same spatial location. They can be lumped into the statement that the structure of the interior of the magma chambers is very uncertain. The error analysis needs broadly to address the effect of this uncertainty on predictions of stresses.

As noted above, tomography has very little input on the depth dependence of viscosity, except to identify regions that are partially molten and thus presumably very invicid. Inferences about the viscosity structure are therefore derived mainly from thermal modeling (which has its own errors).

Assessing Error. Our approach is to first identify a series of 'geophysical observables' and then to estimate their error by Monte-Carlo simulation. Geophysical observables would include the stress, Columb failure function (Deng and Sykes, 1997), vertical uplift, horizontal extension, etc. at places where seismic and/or geodetic measurements are available (or likely to be made in the near future). By Monte-Carlo simulation, we mean that we would run through the complete sequence of tomographic inversion, scaling, reparameterization and elastic (or viscoelastic) calculation multiple times, in such a way to assess the effect of the various types of uncertainties on the observables (i.e. build histograms of probablle ranges of values). At first glance this might seem a rather brute-force technique (and I suppose it is), but on the other hand there is considerable eqperience in the Earthquake Hazard community (with which I have some contact) for using exactly this technique to assess uncertainties in hazard probabilities (e.g. Field and Jacob, 1993). We briefly describe each relevant step.

Tomography. We plan to use the raytrace3d code to re-invert the West (2001) traveltime data for a variety of different points on the trade-off curve of resolution and variance. As this choice is controlled by just one or two parameters, and since a each inversion takes but a few hours on a fast workstation, we can easily prepare a suite of O(100) models.

Scaling. We will conduct a literature survey to assess the amount of variation in scaling laws. Viscosity is likely to be the parameter with the largest uncertainty, as it is largely dependent upon assumptions about the thermal structure of the region. We plan to use some simple 2-d thermal models to make estimates of the probable range of variability of crustal temperature.

Parameterization. We will first perform some initial tests using the Beasy code to quantifty what magnitude and wavelength of parameterization errors are likely to have a significant effect on predictions of the geophysical observables. We will the develop one or more automated regridding strategies that take the scaled tomographic model and produce an input model file compatible with Beasy's requirement that model elements be homogrenous. One reasonable strategy is a three-dimensional analog of replacing a linear gradient with a series of piecewide constant functions (or “steps”): we would first determine isovelocity surfaces within the tomographic model, and then choose elements whose surfaces coincide with them. We would then generate several parameterizations with different steps sizes. We need to pay particular attention to uncertainties introduced by the description of the magma chambers. We plan to assess the differences between several “end-member models”, including magma chambers with a sharp solid/fluid boundary, with transition zones of various thickness and with finite shear modulus throughout.

Reparameterization for the FEVER code is believed to be less of a problem, since the underlying interpolation scheme is more compatible with raytrace3d's.

Modeling of Geophysical Phenomenon

Seismicity. We will examine how the deflation of one of the magma chambers brings regions of the surrounding rock closer to (or farther from) brittle failure (as might be quantified by a Columb failure criterion). Because of the irregular shape of the magma chamber, we expect the stress pattern to significantly depart from the ideal axial-symmetric case. We will compare these predictions with the observed pattern of seismicity following the 1993 and 1998 eruptions (which were well-monitored, both by hyroacoustic means (Dziak et al. 1995,Dziak et al. 1999; Dziak & Fox 1999) and by temporary OBS deployments (Sohn et al. 1999). These sesimicity patterns have several unusual features. For instance, in the year following the 1998 eruption, an area about 5-10 km east of the Axial caldera was unusually seismically active (Sohn et al. 1999) (figure 4A). While this region is well east of the actual eruption, our model shows it to be underlain by a deep extension of the Axial magma chamber. Thus deflation of the magma chamber could possibly lead to loading of this region.

Tidal Loading. Examine how ocean tidal loading effects stresses above magma chamber. Maya Tolstoy, at the 1999 RIDGE workshop in Seattle, presented data that demonstrated that the rate of occurrence of both small, shallow earthquakes and harmonic tremor beneath the Axial caldera is modulated by the ocean tidal cycle. The harmonic tremor has a semidiurnal period, and is probably hydrothermal (water moving through cracks), as contrasted to magmatic in nature. The rate of seismicty is highest at the times of lowest ocean tides. The underlying reason for this modulation is not fully understood. It may be related to the decrease in confining pressure when the weight of the water is decreased. The large difference in the compliance of the magma chamber and the surrounding rock may concentrate stress around the edges of the caldera. Flucuations in pore pressure and in the permeability of the uppermost crust (through changes in normal stress across joint surfaces) may also play a role. Tolstoy et al. (1998) also presented evidence from ocean bottom tiltmeters that the overall magnitude of the deformation associated with tides is much larger, by a factor of 4, than what would be expected from a halfspace model (figure 4C), suggesting that the very compressible material in the magma chamber might be amplifying the tidal signal. We will model this "flexure" of the magma chamber lid and quantitatively assess whether it can explain these diverse phenomena.

Subsidence and Tilting. We will examine how the deflation of one of the magma chambers and diking leads to subsidence, tilting and displacement of the sea floor. Two important uses of geodetic measurements are to determine the net volume loss after an eruption (i.e. estimate the eruption's size) and to measure any widening of the rift zones that may have occurred (i.e. estimate plate-tectonic spreading).

We will address three issues: First, we will first assess whether the 3D model provides significant improvement over the simple "point source in a homogenous halfspace" models that have been applied elsewhere (e.g. Linde et al. 1993). The very significant complexity in the Axial region suggests to us that it will. Second, we will examine the currently available tilt (Anderson et al. 1995; Tolstoy et al. 1998), subsidence data (Fox 1990; Fox 1993; Fox 1999) and extension (Chadwick et al. 1999) data are consistent with the model. The tilt data are fairly limited in scope, with tilt being limited to 9-weeks data from an array of 5 sea floor instruments operated during 1994 (a volcanically quiescent period during which tides were the major signal). The subsidence data, measured using pressure sensors deployed around the caldera, are more voluminous and include the very interesting caldera subsidence that occurred during the 1998 eruption. The extensometer data (figure 4B) includes a very interesting shortening of the north rift of Axial during the 1998 eruption. This is a region that does not contain large accumulations of magma, so the shortening probably represents an elastic response to the deflation of the neighboring Axial magma chamber during the eruption. Chadwich et al. 1999 argue that both the extension and the subsidence can be explained by a Mogi-type model. However, they must place the Mogi pressure source at a depth of 3.8 km, at least 0.7 km shallower than center of the magma chamber, as determined by the tomography (> 4.5 km, see Fig 5). This discrepancy illustrates the need to reconcile the whole range of Axial measurements (including the tomography) in the context of a single model.

Interactions between magma chambers. Examine whether stress interaction between the Axial and Coaxial magmatic systems can plausibly effect the timing of their eruptions. The Coaxial magma chamber is about 15 km from one at Axial, and only about 8 km from the axis of the Axial Volcano's North Rift Zone. One might expect some degree of interaction between the two. Unfortunately, the historical record of eruptions is not long enough to define any sort of eruption recurrence time, and so to interpret the 7 year interval between the 1993 Coaxial and 1998 Axial eruptions. On the other hand, the southward propagation of the 1998 dike is consistent with the 1993 dike having decreased tensional stresses along the North Rift Zone, thus having made northward diking less likely. Our long-term goal here is to develop methodology that will allow the state of stress of the volcano to be tracked over the course of many eruption cycles, and be used to assess the likelihood of future eruptions in various parts of the volcanic system. Such tracking of stress-evolution has proved possible in southern California (Deng & Sykes, 1997) and Turkey (Parsons et al. 2000)

Benefits Of The Proposed Research

This research seeks to build upon the detailed structural information of the volcanic system provided by a 3D tomographic compressional velocity model by using the model to predict geodetic stresses and displacements. The study region, Axial volcano and its immediate vicinity, is one that has attracted intense interest over the years, and which shows prospects of continuing observation (e.g. the Neptune project for a permanent fiber-optic telemetered observatory, http://www.neptune.washington.edu/). Many different types of geophysical data are available now, with features that cannot be explained by models that ignore the strong lateral gradients in material properties related to the presence of magma at shallow depths. Furthermore, more data are likely to become available in the future. This project takes an initial step towards building an integrated model of the volcano, one that has the prospect of allowing a wide variety of data to be modeled; one consistent with the long-term goal of tracking the evolution of the volcano over its next several eruptive cycles.

Management Plan

I (Menke) will be responsible for the timely completion of the project. I, assisted by a GRA, will perform the research.

I am arguably most well-known for my work in tomography and seismic imaging. As this current proposal also relies heavily upon elastic and viscoelastic modeling, I would like to briefly mention that I have considerable experience in that area as well, though less than in tomography and associated with fewer publications. My most recent work is my paper, with Dave Sparks on crustal accretion in Iceland (Menke and Sparks, 1995), which describes a viscous flow / melt extraction model. The calculations were, of course, performed using Spark's convection code, but I was very involved in them. My textbook, Geophysical Theory, with Dallas Abbott (Menke and Abbott, 1989) contains chapters on static stress analysis and viscous deformation, which I wrote. And finally, back in the heyday of plate-bending modeling, I did some work on thin plate flexure (Menke, 1981).

Timetable

This research is expected to take one year.

Dissemination of Results

I will maintain archives of data and preliminary results on myl web site (as I now do for previous studies, see http://www.ldeo.columbia.edu/user/menke). I will make the final RAYTRACE3D, BEASY and FEVER models freely available, so that others can use them. I will present results at scientific national meetings, such as the Fall AGU, and make a best-faith effort to publish them rapidly in peer-reviewed journals.