Single-chamber silicic magma system inferred from shear wave discontinuities of the crust and uppermost mantle, Coso geothermal area, California

Charles K. Wilson, Craig H. Jones, and Hersh J. Gilbert
Department of Geological Sciences and CIRES, University of Colorado, Boulder, Colorado, USA

Received 29 January 2002; revised 21 August 2002; accepted 27 January 2003; published 1 May 2003.

[1] Analysis of seismograms from teleseismic rays traversing the Coso geothermal area near Ridgecrest, California, suggests the geothermal system lies over a single shallow magma reservoir (~5 km below the surface) that also plays a crucial role in the local change in deformation style from areas to the north and west. The character of the magma reservoir and the absence of a lower crustal magma reservoir is inferred from three crustal P-to-S conversions observed using receiver function analysis: (1) A high-amplitude, shallow, negative arrival, Ps-P time of 0.7–0.9 s (3–5 km below sea level (bsl)), (2) a moderate amplitude, positive conversion, Ps-P time of 2.1–2.5 s (14–17 km bsl), and (3) the Moho conversion, Ps-P time of 4.0–4.2 s (30–32 km bsl). Observations of Moho converted arrivals indicate that the interface is mostly flat and uncomplicated throughout the study area, while the midcrustal conversion is laterally variable in amplitude and depth. The absence of the large negative amplitude conversion on waveforms recorded at stations outside the geothermal area strongly suggests that the feature lies only underneath the modern geothermal area. In addition, rays sampling the shallow converter also contain later arrivals with retrograde moveout consistent with an origin as reverberations above the conversion. Receiver functions calculated from synthetic data using a single isotropic layer over a half-space indicates that the shear velocity decreases by 30% across the interface (V\textsubscript{s1} = 2.6 km/s; V\textsubscript{s2} = 1.8 km/s; layer one thickness 4.9 km), further supporting the presence of shallow magma.

INDEX TERMS: 7205 Seismology: Continental crust (1242); 7299 Seismology: General or miscellaneous; 8015 Structural Geology: Local crustal structure; 8109 Tectonophysics: Continental tectonics—extensional (0905); 8434 Volcanology: Magma migration; KEYWORDS: seismic arrays, receiver function analysis, crustal seismology, silicic volcanic center, basin and range, magma genesis


1. Introduction

[2] Volcanic plumbing in the lower crust remains poorly understood. End-member hypotheses include both single chamber magma systems from which a full spectrum of melts are produced through magma mixing and gravitationally induced stratification and multichamber systems that allow melt compositional changes to occur through differentiation within several crustal reservoirs. Geophysical studies of volcanic systems have provided some insight into this problem [e.g., Reasenberg et al., 1980; Lutter et al., 1995; Masturyono et al., 2001; Weiland et al., 1995; Plouff and Isherwood, 1980], but often results are ambiguous because of vertical tradeoffs in structures derived from gravity and teleseismic travel time measurements, inherent difficulties produced by low-velocity zones, and the absence of deep crustal earthquakes in the volcanic centers. Geochemical efforts also suffer from ambiguities, which make it difficult to discern between the various models of reservoir development [Duffield et al., 1980; Bacon et al., 1981, 1980; Manley and Bacon, 2000; Verplanck et al., 1999].

[3] One of the largest geothermal fields related to young silicic volcanic centers in the Basin and Range, the Coso geothermal field has been extensively studied geochemically [Adams et al., 2000; Duffield et al., 1980; Bacon et al., 1981; Manley and Bacon, 2000], structurally [Duffield et al., 1980; Roquemore, 1980; Whitmarsh, 1998], gravimetrically [Plouff and Isherwood, 1980], and seismologically [Reasenberg et al., 1980; Young and Ward, 1980; Lees and Wu, 1999; Wu and Lees, 1999; Lees and Wu, 2000, Lees, 2001]. Geochemical studies have established possible compositions and source depths for the extruded melts while the geophysical studies have been helpful in outlining the depth and character of the top of the upper crustal

Copyright 2003 by the American Geophysical Union.
0148-0227/03/2002JB001798S09.00
reservoir and geothermal production area. However, like other recently studied volcanic systems, the connection of the upper crustal magma reservoir to its deeper origins remains poorly resolved. To explore the whole volcanic plumbing under the Coso geothermal area and its possible relationship to regional tectonics, the lithospheric structure of the area was studied using receiver functions computed from seismograms recorded by a group of high-density seismic arrays deployed between November 1998 and May 2000.

2. Geologic and Tectonic Setting

The Coso geothermal field near Ridgecrest, California, occupies the corner between three tectonic regions: the central Basin and Range, the Sierra Nevada block, and the Mojave Desert. The geothermal field is in the western part of the Eastern California Shear Zone [e.g., Dokka and Travis, 1990]. Large-scale west directed normal faulting characterizes this part of the Basin and Range [e.g., Wernecke, 1985] and may root into a low-angle deformation zone that extends under the Sierra. This shear zone extends under the Sierra and Basin and Range to the north [Jones and Phinney, 1998] and possibly northeast [Zhou and Phinney, 2000] of the Coso region. Jones [1987] suggested that the Coso area acts as an “accommodation” terrane, absorbing differences between northwest directed extension to the north and west-southwest oriented strike-slip faulting along the Garlock Fault to the south. This supports a change in deformation style from that inferred to the north by Jones and Phinney [1998] to something different near the Coso geothermal area.

Two distinct episodes characterize magmatic production in the Coso area over the past four million years. The first episode erupted a broad spectrum of volcanic rocks (basalt, andesite, dacite, and rhyolite) from 4 Ma to 2.5 Ma. This was followed by a period of intense bimodal magmatism (Figure 1) from 1 Ma to the present. This produced 39 high-silica rhyolite domes with basalt flows along the periphery of the rhyolitic volcanism [Duffield et al., 1980; Bacon et al., 1981]. Sugarloaf Mountain, emplaced ~40,000 years ago, is the youngest dated dome and is the latest surface expression of a partially molten silicic magmatic reservoir inferred to be present between 5 and 20 km depth beneath the modern geothermal field [Adams et al., 2000; Duffield et al., 1980; Reasenberg et al., 1980; Bacon et al., 1981; Young and Ward, 1980]. Surface mapping of hydrothermally altered deposits supports the existence of the modern geothermal field for at least the last 300,000 years [Adams et al., 2000]. Duffield et al. [1980] suggested that the heat for this system has been recharged through mantle derived basaltic magmas typical of recent Basin and Range volcanic development.

The upper few kilometers of the magmatic system contains dramatic variations in seismic properties [Lees and Wu, 1999; Wu and Lees, 1999; Lees and Wu, 2000; Malin, 1994; C. W. Caruso et al., A seismic transect of the western Coso Range, eastern central California, unpublished manuscript, 1994]. Below it lies a sharp, shallow brittle-ductile transition [Lees, 2001] and reversed polarity reflection about 5 km below the surface (C.W. Caruso et al., A seismic transect of the western Coso Range, eastern central California, unpublished manuscript, 1994) Unfortunately, these recent studies do not penetrate the top of the magmatic system; deeper levels were sampled using teleseismic P wave tomography by Reasenberg et al. [1980], who found low velocities in the crust but had limited resolution due to station source geometries available. Young and Ward [1980] found attenuation of teleseismic arrivals near the geothermal field. Other surveys using local earthquakes were focused to the south and found little variation in this region, though a paucity of ray paths could explain the absence of identified anomalies [Walck and Clayton, 1987; Walck, 1988; Sanders et al., 1988]. Thus questions remain about the nature of the discontinuity near 5 km depth (is it purely hydrothermal or does it represent the top of a magma chamber?) and the underlying magmatic system.

3. Data and Method

We recorded ~220 Gb of 40 samples per second, three-component seismograms in and around the Coso Geothermal Area from November 1998 to May 2000; most of the data is available from the IRIS Data Management Center (http://www.iris.washington.edu). With over 150 sites occupying an area of ~2000 km², this is one of the densest portable, passive seismic deployments to date (Figure 1). In an effort to support further use of high-density seismic arrays for lithospheric study, we give a detailed explanation of the experiment design and data processing methods used in this study.

3.1. Array Design and Placement

The employment of arrays to improve crustal imaging has been widely used with much success in multichannel reflection processing [e.g., Yilmaz, 1997]. Array processing enhances the desired signal while suppressing scattered energy from short wavelength topographic and subsurface variations [Abers, 1998; McNamara and Owens, 1993; Jones and Phinney, 1998]. Our arrays consisted of 5–8 short-period sensors (1 or 2 Hz free period: Mark Products L22 or L4c, and/or Teledyne Geotech S-13) spaced 500 m apart and arranged into two orthogonal lines with, when possible, on bedrock insofar as was possible; in most cases, the arrays are on Mesozoic plutonic rocks or a thin (<100 m) veneer of Quaternary sediments over Mesozoic bedrock.
the edges of the Airport Lake basin, and POR, which, though on the footwall of the Sierra Nevada frontal fault, is over an unknown and variable amount of sediment.

### 3.2. Event Selection

An empirical relationship involving magnitude ($\Omega$, maximum of $m_b$ and $M_s$), depth ($\Gamma$, km), and distance ($\Delta$, in degrees) developed by Jones and Phinney [1998] was used to select teleseismic events for examination:

$$\Omega - 0.0055*\Delta + 0.0016*\Gamma \geq 4.24.$$  \hspace{1cm} (1)

From these, 244 events, ranging from 29° to 95° distant, were chosen according to signal-to-noise ratio estimates for

**Figure 1.** Shaded relief map with geology centered on Coso geothermal area shown with a western United States tectonic map to the right. The red box on the tectonic map indicates the location of the study area. Geologic data used in this map can be found at http://geomaps.geo.ukans.edu/coso.html.
the $P$ arrival with consideration given to the overall backazimuthal distribution of events (Figure 2). Traces with spurious noise from equipment failures or other causes were dropped from processing.

### 3.3. Data Preparation

Careful visual checking of each event-array pair insured proper timing and quality of traces on all components. Shifting and stacking the recorded waveforms according to predicted moveout of the direct $P$ phase helped in checking station timing for a given event. In a very few cases, a station obviously misstacked due to timing errors (>0.1 s) was removed for that event. The theoretical instrument response was removed from each trace and replaced with the theoretical response of a Mark Products L4c seismometer (free period is 1 s, damping parameter is 0.7). After response correction, beams were formed for each event-array pair for the vertical, north, and east components by time shifting seismograms to remove plane wave moveout of the $P$ phase [Abers, 1998; McNamara and Owens, 1993; Langston and Hammer, 2001]. We calculated the slowness for each event using the IASPEI91 velocity model [Kennett and Engdahl, 1991] and locations reported in the Preliminary Determination of Epicenters (PDE catalog) (as archived at http://iris.washington.edu/). A surface $P$ velocity of 5.5 km/s was used to correct the theoretical moveout for topography. Other near-surface velocities were tested but the resulting beams showed little variation probably due to the small incidence angle of the incoming arrivals. Rotations of horizontal component beamed seismograms into the radial and transverse system followed beam forming because both horizontal components weren’t always available at each site. Before further processing, the beams were band-pass filtered in the frequency range of 0.3–6 Hz to remove microseismic noise and high-frequency background noise present at some arrays.

### 3.4. Receiver Function Calculation and Compositing

The goal of receiver function analysis is to enhance the arrivals of receiver side $P$-to-$S$ converted phases (Figure 3) [Burdick and Langston, 1977]. This is done through deconvolution of the vertical from the radial component seismogram. The result is the receiver function, which represents arrivals of various $P_dS$ conversions after the direct $P$ wave. In this paper we follow the convention that all letters following the initial phase are upper case if downgoing and lowercase if upgoing and the subscripted letters denote the new phase initiation point with the letter $d$ symbolizing some arbitrary velocity discontinuity (Figure 3). Thus a Moho conversion is $P_dP_{dS}$ and a reverberation of an upgoing $P$ to a downgoing $P$ to an $S$ reflected from the Moho would be $PpP_{dS}$. The amplitude of the converted arrival is a function of the impedance contrast across an interface and the incidence angle of the incoming wave [Langston, 1977]. A conversion from the top of a low-velocity body will produce a conversion reversed in polarity from the impinging wavelet. Particle motion of $P_dS$ conversions from flat lying planar isotropic structures will be radially polarized, therefore a conversion from a discontinuity should be consistent for all back azimuths for the same incidence angle.

We calculated radial receiver functions from the filtered, beamed seismograms. Several techniques yielded similar results: a standard frequency domain Oldenburg deconvolution [Oldenburg, 1981], a least squares time domain deconvolution [e.g., Abers et al., 1995], and an iterative cross-correlation deconvolution [Ligorria and...
Ammon, 1999]. Some variations between the results are most likely related to the damping parameters chosen for the particular deconvolution method [Ammon, 1992]. We use an iterative method relying on a cross correlation of the radial and vertical component seismograms [Ligorria and Ammon, 1999]. A spike on the evolving receiver function is inserted at the location of the largest amplitude of the cross correlation with the amplitude of the spike scaled by the inverse of the vertical autocorrelation. After each iteration, the receiver function is convolved with the vertical component seismogram. The result is subtracted from the radial to remove radial energy already accounted for by the receiver function. The modified radial is then used for subsequent iterations. The convolution of the receiver function with the vertical seismogram is compared to the appropriate filtered radial seismogram to calculate the variance reduction. Once the variance reduction improvement between iterations falls below a threshold or the number of iterations reaches a predetermined limit, the procedure is terminated. Receiver functions reducing the variance by greater than 70% were used in the final analysis.

[14] To compare and stack events from different distances and back-azimuths, we trace $P_s$ rays under the region to 200 km depth using the same velocity structure as Jones and Phinney [1998] (Figure 4). A thick gradational Moho instead of a sharp boundary between crust and mantle velocities reduces errors in amplitude and depth from improper placement of the Moho in an a priori model. Ray tracing allows the projection of the conversions to a pseudodepth, making single array back-azimuthal stacks and common conversion point stacks possible. In addition, we correct receiver function amplitudes to those of a fixed incidence angle [Jones and Phinney, 1998]. Rescaling of conversion amplitudes prevents bias when stacking over varying incidence angles and permits interpretation of arrival amplitude as directly proportional to the $S$ wave impedance contrast across an interface.

### 3.5. Common Conversion Point Stacking

[15] To study the lateral variations in subsurface features and further mitigate effects from unwanted energy, we stack the receiver functions (e.g., Figure 5) into common conversion point (CCP) bins (Figure 6). Significant amounts of energy on receiver functions can result from signal-generated noise: for instance, scattering from surface basins [Levander and Hill, 1985] and nonplanar interfaces [e.g., Clouser and Langston, 1995]. Stacking data from multiple arrays together further diminishes signal-generated noise and scattered energy that may be coherent at one array, but commonly stacks down when combining data from multiple arrays.

[16] The geographic stacking procedures used here largely follow those of Dueker and Sheehan [1998] with...
modifications. $P_d$ arrival piercing points, which have been found using the ray tracing techniques described above, are binned using a 3-D grid of sample points spaced 3 km apart horizontally and 0.25 km apart in depth within a 47 (east-west) by 43 (north-south) by 65 (vertical) km volume. The piercing points within a $9^2 \times 9$ km$^2$ area centered on each sample point are stacked at that point (Figure 6). Array elevation corrections were made before common conversion point stacking based on the predicted angle of incidence, the station elevation, and near-surface velocity information from recent active source studies [Unruh et al., 2000, 2001a, 2001b; Pullammanappallil et al., 2001]. Changing the size of the bins, or weighting rays differently depending on their distance from the bin center point, does not greatly influence our results. Bootstrap resampling with replacement within individual CCP bins, performed in order to estimate the error associated with that trace [Efron and Tibshirani, 1986], helps assure that a few errant receiver functions do not control the results. We calculated one hundred bootstrap realizations of each stack and present their mean and standard deviation for each bin to give an idea of the reliability of the results for each bin.

[17] Images produced by common conversion point stacking may suffer from effects of velocity variations that could influence interpretation if unrecognized. All the depths presented are calculated by tracing rays through a one-dimensional structure as discussed previously. Lateral variations in seismic velocities will cause features to be mapped to erroneously deep depths below low $S$ velocity bodies and shallow depths beneath high $S$ velocity bodies. In extreme cases, conversions from a single point to differ-

Figure 5. Back-azimuthally binned receiver functions from four arrays shown on the map in Figure 1. The receiver functions have been projected to depth using the CCP stacking velocity model shown in Figure 4. The depth is plotted on the horizontal axis in kilometers below sea level. The back-azimuth bin is shown on the left vertical axis. The right vertical axis indicates the number of receiver functions in each bin. This attribute is also indicated by the change in saturation of the fill for each binned receiver function. These plots demonstrate the rapid change of the UCN depth within individual arrays. (left) Step-like east-west variations from arrays CGT and CDR. (right) Smooth north-south variations of the UCN depth at arrays JSH and SSL.
4. Seismological Observations

[18] Analysis of the receiver functions reveals a wealth of converted energy from the crust under this region. We identify three principal discontinuities that help to illuminate the tectonic features of the crust in this region (Figures 5 and 7):

[19] 1. The upper crustal negative (UCN) discontinuity is a shallow negative conversion characterized by a high-amplitude swing at 4–5 km bsl near the geothermal field. The discontinuity appears to dip to the north along its eastern edge. The depth and amplitude of this converter vary over short horizontal distances to the south and west of the geothermal field.

[20] 2. The Moho discontinuity is a prominent positive arrival that runs throughout the study area at ~31 km bsl. Previous studies from this area placed the Moho at similar depths [Fliedner et al., 2000; Jones and Phinney, 1998].

[21] 3. The midcrustal positive (MCP) discontinuity is a positive discontinuity characterized by a high-amplitude pulse near 15 km bsl. Under the upper crustal negative, the amplitude of the arrival decreases as the depth increases by 2–3 km.

[22] The following discussion details important changes in these conversions found on CCP stacked cross sections as well as single array back-azimuth stacks. We provide the CCP sections and back-azimuth stacks pertinent for the following observations.

4.1. Upper Crustal Negative Location and Character

[23] The shallow negative arrival observed on the individual array back-azimuth stacks (Figure 5) and the cross sections through the center of the geothermal field (Figure 7; cross sections located on Figure 6), appears strongest in the region between 4 km east to 10 km west and 5 km north to 10 km south of the coordinate center at Coso Hot Springs (Figure 1). The converter resides near 3 to 4 km depth below sea level in this area (Figure 7). The large amplitude of the UCN arrival suggests a dramatic decrease in shear velocity below the interface. The depth of this low-velocity zone is consistent with previous geophysical studies from the Coso geothermal area [Lees and Wu, 1999; Wu and Lees, 1999; Lees and Wu, 2000; Lees, 2001; Malin, 1994; Unruh et al., 2000, 2001a, 2001b; Pullammanappallil et al., 2001; C. W. Caruso et al., A seismic transect of the western Coso Range, eastern central California, unpublished manuscript, 1994].

[24] The depth of the UCN varies over short distances at the southern and western edges of its core area (Figure 8). The single array back-azimuth stacks from JSH, SSL, CDR, and CGT show rapid variations in the depth of the UCN with back-azimuth (Figure 5). JSH shows a deepening to the north and west from depths of 7 km, while a few kilometers to the north, SSL's arrivals shallower to 5 km. Assuming these shallow arrivals indicate the location of the shallow discontinuity beneath each array, the depth variation between arrays may be explained by a grossly east-west trending fault with 2 km of apparent vertical offset. The deeper arrivals to the north and west observed at JSH can be explained as a diffraction tail generated by arrivals encountering the fault from different directions. The deepest arrivals at SSL are sampling the offset UCN from beneath the JSH array and to the east of JSH. This is probably the high-angle fault inferred from geologic mapping to trend about north-west between SSL and JSH [Roquemore, 1980; Bacon et al., 1980; Whitmarsh, 1998; also see http://geomaps.geo.ukans.edu/coso.html].

[25] To the west, CGT and CDR image a dip in the western boundary of the converter. For eastern back-azimuths, both arrays show a negative arrival from a depth near 6 km. UCN arrivals from the west indicate a deeper origin near 9 to 10 km depth. This could point to the presence of a north-south trending fault with a 3 to 4 kilometer offset or a strong dip of the conversion to the west. This edge of Rose Valley is the site of significant normal faulting [Roquemore, 1980; Bacon et al., 1980; Whitmarsh, 1998; also see http://geomaps.
4.2. Complications of the Moho and Midcrustal Positive Arrivals

[26] Detailed study of the CCP stacked sections reveals large short-wavelength variations in amplitudes of the Moho and the MCP conversions. In cross section 1 (Figure 7), the location of low-amplitude and distorted Moho arrivals is under the stronger parts of the shallow negative anomaly. The MCP arrival shows similar amplitude variations, which is particularly clear in cross section 1 (Figure 7, top). In addition, the MCP appears to be depressed by several kilometers under the UCN.
character and extent of the UCN conversion from CCP stacks. Colored blocks are the amplitude of the conversion; converter depth is contoured in kilometers below sea level for the regional velocity structure (depths shallow by 1.5 km using the UCN velocity structure). Rays penetrating the UCN (for purposes of stacking, e.g., Figure 9) are in yellow; others are blue and are plotted at 5 km depth. Note increasing depth and decreasing amplitude to the south and west. Green line separates rhyolite-dominant interior from basalt-dominant exterior. Coso Hot Springs is at the corner of the “L” of the CHS array.

[27] The presence of a strong UCN conversion above lateral changes in the deeper converters suggests either that the UCN causes these changes by attenuating and/or delaying arrivals from greater depth, or that the UCN is part of a larger feature extending to the Moho. The former seems most plausible: an area of high shear wave attenuation associated with the shallow negative conversion could attenuate converted rays from deeper discontinuities such as the Moho or the MCP. Converted energy from these discontinuities traversing the low-velocity region would appear to be from greater depths due to conversion misprojection related to the low-velocity zone. However, these surmises do not inform us of the depth extent of anomalously slow crust.

[28] To consider the effect of upper crustal elastic parameter variations on deeper conversions, the complete CCP sections were remade excluding the rays piercing the high-amplitude sections of the UCN (Figure 7, middle). In this image, the Moho arrival appears consistent and at a nearly constant depth across the images. In addition, the MCP arrival has less topography and stronger amplitude beneath the UCN in the new images. The sections stacked only with rays piercing the UCN (Figure 7, bottom) show several high-amplitude midcrustal arrivals that will be discussed in detail in the next section, but the Moho and MCP arrivals are weak or even nonexistent at the depths seen in Figure 7 (middle). These deeper conversions might be present at greater model depths (e.g., the Moho might be the positive anomaly ~42–44 km model depth); this would be consistent with large velocity pull down from a major low-velocity zone. These observations relate the amplitude and topography of deeper conversions to low velocities and moderate attenuation of the region at and immediately beneath the UCN.

4.3. Post-UCN Arrivals

[29] The restacked CCP images demonstrate the presence of a high-amplitude negative arrival from near 23 km depth and a low-amplitude positive arrival from ~29 km depth on the cross sections using only UCN rays (Figure 7, bottom). The arrivals are noticeably absent on the cross sections where UCN rays were excluded (Figure 7, middle). An important attribute of these arrivals is their confinement beneath the high-amplitude portions of the UCN.

[30] To determine the nature of these localized post-UCN arrivals, the receiver functions from rays sampling the UCN and rays not sampling the UCN were stacked in ray parameter bins to examine variations in arrival time with changing event distance (Figure 9). Commonly known as moveout or tau-p analysis, plotting receiver functions in this manner allows the easy identification of reflected phases because of their reverse or retrograde moveout [Gurrola et al., 1994]. The solid lines indicate the theoretical moveout curves for a conversion from a discontinuity shown at the depth indicated to the right at the top of the curve. The most prominent arrivals are the high-amplitude, shallow, negative arrivals from 4.9 km below the surface on the UCN receiver function plot and the strong, consistent Moho on the plot with rays not sampling the UCN. Other arrivals include the PPUCN reverberation and the PpPUCN + PpSUCN reverberation off a discontinuity at 4.9 km below the surface marked by the long dashed and short dashed curves, respectively (Figure 9; see Figure 3 for illustration of the various back-scattered and forward scattered phases). The retrograde moveout demonstrated by the theoretical reverberation curves make the reflected arrivals easy to distinguish from direct phases. Both the negative arrival at 23 km depth on the CCP stacks and the positive arrival near 29 km depth (Figure 7, bottom) match predicted times, polarity, and retrograde moveout of reverberations from the top of the low-velocity zone at the UCN.

[31] Receiver function modeling using synthetic data generated by the technique of Keith and Crampin [1977] helped to further constrain the shallow velocity structure within the geothermal field. All receiver functions from rays piercing the UCN region described in the beginning of section 4.1 were stacked to create a mean representative receiver function for the UCN region. Receiver functions calculated from synthetic data were compared to the mean UCN receiver function until the best fit was found. The best fit criteria were arrival time, amplitude, and pulse shape. The model space was thoroughly explored, with particular attention paid to the layer 1 P and S velocities, the S velocity contrast across the UCN interface, and the density contrast across the interface (Table 1 indicates possible variations of these model parameters). To fit the amplitude of the various conversions and reverberations, the two-layer isotropic velocity model requires a decrease in shear wave velocity...
across the UCN of 30% (+13/−3%) (Figure 10 and Table 1). The \( PpP_{UCN} \) arrival is sensitive to density contrasts making it possible to constrain the density contrast across the UCN to be \(-0.15 \text{ g/cm}^3\) (−0.05/+0.15 g/cm\(^3\)). The resulting synthetic receiver function fits the observed mean receiver function for the primary \( P \)-to-\( S \) conversion as well as the \( PpPs_{UCN} \) arrival but the amplitude for the \( PsPs_{UCN} + PpS_{UCN} \) arrival is larger than that observed. The loss in amplitude is most likely due to the limited lateral extent of the UCN and its effect on the \( PsPs_{UCN} + PpS_{UCN} \) arrival (Figure 10).

Figure 9. Receiver functions stacked according to ray parameter (with averaging over 0.006 s/km) for (top) rays sampling the UCN and (bottom) rays not sampling the UCN. The horizontal axis represents time after the \( P \) arrival. The right vertical axis represents the number of receiver functions used within each ray parameter bin. The solid curves show calculated moveout curves of converted arrivals from depths plotted at the top of the curves. The long dashed line and short dashed line represent \( PpPs \) reverberation \( PsPs + PpSs \) reverberation from 4.9 km depth, respectively. The curves in the top plot were calculated using the velocity model shown as a dashed line in Figure 4. The curves of the bottom plot were calculated using the standard CCP stacking velocity (Figure 4).

[32] Reflectivity based synthetic modeling [Kennett, 1983], which permits variations in \( Q \), produces slightly better pulse shape matching of the forward scattered converted arrival, but the required density contrast necessary to fit the \( PpPs \) reverberation is unrealistic. The predicted amplitude of the \( PsPs + PpSs \) arrival is still a factor of 2 too large. Therefore we believe that although attenuation in the region above the UCN may be important, it may be difficult to distinguish from other effects such as destructive interference of multiply scattered waves generated within the low-velocity region.

[33] The absence of significant pull-up or push-down on the Moho and possibly MCP conversions permits us to address the thickness of the low-velocity zone beneath the UCN conversion. A simple geologic structure consistent with previous teleseismic imaging [Reasenberg et al., 1980] would extend the low-velocity zone down to 15 km bsl without change in velocity. The base of this zone would
then be a positive increase compatible with the presence of the MCP but probably demanding much greater amplitude. It would also move the expected Moho conversion arrival time to almost 7.5 s after the direct $P$ arrival, where no arrival is seen (Figure 9). Finally, enough of the rays not penetrating the UCN do extend under it enough to have their Moho conversion delayed were this structure correct; the Moho does not show such pullup (Figure 7, middle). Unless a compensatory high-velocity body directly underlies this low-velocity zone with a geometry preventing detection, these observations limit the vertical extent of a constant velocity low-velocity zone to about 10 km.

The core UCN region is bounded on the south and west by sharp increases in the converter depth, which we earlier related to faulting (section 4.1, Figure 5). These terminations are not far from the edges of the recognized geothermal field. Structural termination of the UCN strongly indicates that upper crustal geothermal systems (and possibly magmatic systems) can be controlled by tectonic boundaries and do not merely taper out laterally. These observations may also support previous assertions that the buoyancy of the magma body may be responsible for arcuate structural features observed in the area [Duffield et al., 1980].

### Table 1. Parameters Used in Receiver Function Modeling

<table>
<thead>
<tr>
<th>Layer</th>
<th>Model 1</th>
<th>Model 2</th>
<th>Model 3</th>
</tr>
</thead>
<tbody>
<tr>
<td>$V_p$, km/s</td>
<td>4.95</td>
<td>4.50</td>
<td>4.90</td>
</tr>
<tr>
<td>$V_s$, km/s</td>
<td>2.55</td>
<td>1.8</td>
<td>2.60</td>
</tr>
<tr>
<td>Density, g/cm$^3$</td>
<td>2.75</td>
<td>2.7</td>
<td>2.7</td>
</tr>
<tr>
<td>Thickness, km</td>
<td>4.9</td>
<td>-</td>
<td>4.9</td>
</tr>
</tbody>
</table>

Model 1 is the preferred fit and the result is shown in Figure 10. Models 2 and 3 have been provided to show the sensitivity of the modeling to various parameters. Model 2 fits the upper error estimates of the mean receiver function, while model 3 fits the lower error estimates.

5. Discussion

5.1. Geography of the UCN

The UCN conversion is the strongest crustal conversion in the area and is tightly tied to the geothermal field and recent silicic volcanism. Thus this converter is a product of Quaternary volcanism and provides important clues to the structure of the Coso volcanic center. The discontinuity is highly localized, with a center just south and west of the Coso Hot Springs area (Figure 8). The amplitude is greatest within a 14 km by 15 km square area (~210 km$^2$) where the converter is about 5 km below the surface when using the reverberation velocity structure of Figure 4 (Figures 5, 7, 9, and 10). The boundaries on this region are only defined within 3–5 km because of spatial averaging of the stacking techniques. Ray coverage is particularly poor to the southeast (Figure 8), and the largest errors in the UCN boundaries may be found in this region (Figure 6).

The core UCN region is bounded on the south and west by sharp increases in the converter depth, which we earlier related to faulting (section 4.1, Figure 5). These terminations are not far from the edges of the recognized geothermal field. Structural termination of the UCN strongly indicates that upper crustal geothermal systems (and possibly magmatic systems) can be controlled by tectonic boundaries and do not merely taper out laterally. These observations may also support previous assertions that the buoyancy of the magma body may be responsible for arcuate structural features observed in the area [Duffield et al., 1980].

![Figure 10](image.png)

Figure 10. Synthetic receiver function calculated from model 1 (Table 1) plotted with the mean receiver function for the rays sampling the UCN anomaly. The dashed lines represent the standard deviation associated with the mean receiver function. The $P_{ds}$ and $P_{PP}$ arrivals are fit nicely, but the synthetic amplitude exceeds the observed amplitude for the $P_{PP} + P_{SS}$ arrival. The diagram included in the waveform window demonstrates one contribution to the observed decrease in amplitude. The loss of expected energy is due to the limited area where the $P_{SS}$ ray can be generated.
[37] As noted before, existing geophysical work provides few constraints on the magmatic system beneath the UCN. The low-velocity body could either be a hydrothermal system, possibly highly overpressured, well above magma, or it could be magma itself. The rhyolitic volcanism and petrologic relationships in the volcanic field require a crustal magma chamber [Manley and Bacon, 2000], so if there is not another candidate magma chamber, we can proceed to interpret the characteristics of the upper crustal low-velocity zone for magma present. We first consider the possibility of deeper magma chambers beneath the observed low-velocity zone, and then use geophysical observations to constrain the bulk properties of the crust under the geothermal area. Finally, we address the implications of our inferred structure for deformation of the crust.

5.2. Absence of Lower Crustal Magma Bodies

Some geophysical studies of other volcanic systems support the presence of a lower crustal reservoir [Lutter et al., 1995; Weiland et al., 1995; Steck et al., 1998] or melt-filled basaltic feeder system extending to a mantle source [Masturyono et al., 2001]. Any candidate magma bodies within the crust should have low S velocities and thus delay any deeper Ps conversions and generate Ps conversions of their own, particularly at the top where a velocity decrease should exist.

A lower crustal magma pool would certainly produce large negative conversions comparable to the UCN on both UCN-penetrating rays and rays from adjacent arrays that undershoot the UCN. The only plausible candidate is between CHS and BPT at a depth of 25 km (cross section 1, Figure 7, middle); the amplitude of this anomaly is a quarter of the UCN, and the anomaly is only observed from a single array (BPT), although it might be delayed and present in the CHS gathers. We view this as a very weak candidate for a lower crustal magma chamber. Otherwise, the absence of strong conversions and the complications in Moho depth such as a velocity perturbation should introduce to the CCP sections make such a body unlikely under Coso. A pervasive, melt-filled feeder dike system might not produce strong and coherent Ps conversions, but the absence of velocity pullup or pushdown artifacts on the Moho image make this an unlikely scenario. Instead, the upper crustal reservoir may be fed either through discrete injections of basaltic melt from the mantle or by small dikes. Basaltic injections through these mechanisms would provide heat and volatiles to the previously stratified reservoir while leaving the seismic velocities in the lower crust sufficiently high that artifacts would not be produced on our images of the Moho. Input of volatiles and heat to the upper crust reservoir eventually forces eruption of high silica magma from the top of the reservoir, producing one of the many high silica rhyolite domes. Injections not trapped by the upper crustal reservoir would be extruded outside the area of rhyolite domes, producing the bimodal magmatism and allowing the main geothermal field to remain mostly free of basalt [Manley and Bacon, 2000].

To establish the limits of the width of melt-filled basaltic dikes feeding the upper crustal magma system without significantly reducing lower crustal velocities, we consider vertical dikes (or vertically elongated magma chambers) of varying widths and their impact on the timing of the Moho Ps arrival. Assuming seismic waves are sensitive only to structural features longer than a quarter wavelength, and noting that teleseismic energy recorded by our short-period instruments is peaked near 1 Hz, a dike 500 m wide could remain undetected by incoming waves traveling at velocities as low as 2 km/s (a more normal lower crustal P velocity of 6 km/s might not detect a dike 1.5 km wide). In addition, if this main dike were divided into smaller dikes and distributed throughout the lower crust beneath the UCN, then the effect on individual teleseismic arrivals would be even less. Thus we suggest that any melt-filled dikes currently present under the upper crustal magma body can be no more than a few hundred meters wide and must be distributed irregularly through the area.

[41] Attenuation in the crust can also be used to explore for melts and fluids. We observe attenuation of Ps conversions traversing the UCN: Moho and MCP converters are noticeably lower in amplitude for these rays (Figures 7 and 9). Previously, Young and Ward [1980] and Lees [2001] inferred the presence of a high-attenuation region in the shallow crust above the depth of the UCN (<5 km depth), presumably a reflection of severe hydrothermal activity. Young and Ward [1980] also inferred some attenuation from the mid to lower crust (12–20 km depth) within the Coso area. If most of the attenuation is from the top 5 km, as Young and Ward [1980] suggest, then the UCN is an even greater velocity contrast than in our velocity structures, and the presence of the strong PpPs reverberation above the UCN would be difficult to explain. In contrast, an absence of attenuation within the low-velocity zone below the UCN should produce large amplitude reverberations from within that layer. Attempts to model such reverberations with synthetic seismograms lacking attenuation produced large-amplitude arrivals not visible on our seismograms. We suggest that the bulk of the attenuation and Young and Ward [1980] observe is from between about 5 and 15 km depth and that this is probably compatible with the limited vertical resolution of the teleseismic observations used by Young and Ward [1980].

5.3. Melt in the Upper Crustal Magma Chamber

From our analysis above, it seems likely that the body under the UCN is magma, since there is no deeper candidate within the crust; some amount of hydrothermal fluids immediately above the UCN must exist to drive the geothermal field. Nakajima et al. [2001] showed a relationship between $V_p/V_S$ and $V_p$ for crustal rocks with water and melt filled cracks of different aspect ratios. A $V_p/V_S$ and $V_p$ similar to our modeled values for layer 1 within the geothermal field ($V_p/V_S = 1.94$; $V_p = 4.9$ km/s; Figure 4) would predict a system with a crack aspect ratio between 0.01 and 0.001 and $\sim$1–2% hydrothermal fluid. Estimates of the layer 2 P velocity are ambiguous because of low sensitivity of the arrival amplitudes to the layer 2 P velocity. Our only control on sub-UCN P velocities is the forward modeling used in matching the Moho arrival on the moveout plots for rays sampling the UCN; this is consistent with previous P wave tomographic studies in this region [Reasenberg et al., 1980]. Velocities directly
beneath the UCN ($V_p/V_s = 2.5$; $V_p = 4.5$) would indicate 1.5–5% melt for a crack aspect ratio between 0.01 and 0.001. If, in fact, the $P$ velocity is also much lower just beneath the UCN, both the percentage of melt and aspect ratio could be much higher.

[43] The gravity signature of the magma reservoir further constrains the character of the magma chamber. Plouff and Isherwood [1980] previously noted the ~10 mGal anomaly associated with the geothermal field. Approximating a magma body as an infinite Bouguer slab, we can determine the minimum thickness of a melt column contained within the region beneath the UCN. Assuming a density of 2700 kg/m$^3$ for the average crustal density, crystal free rhyolite melt would create a density contrast of 400 kg/m$^3$; a crystal free basalt melt would have little or no density contrast [Bergantz and Dawes, 1994; Bacon et al., 1980]. Thus a column of rhyolitic magma represents the minimum amount of magma capable of producing the observed gravity anomaly. The minimum thickness of melt is 0.6 km. If the melt percentage beneath the UCN is no higher than 5%, as suggested by seismic velocities, a column of 5% melt stretching to 17.5 km bsl would produce the observed gravity anomaly. If the melt percentage decays with depth or is compositionally zoned to more mafic compositions with increasing depth; it would be less if a considerable part of the gravity anomaly is caused by hydrothermal alteration of granite in the top few kilometers. Larger melt percentages may be present for more mafic melts. In contrast, if the density contrast of 0.15 Mg/m$^3$ inferred from fitting of reverberations is used, the melt percentage just under the UCN is 30–35%, and such a region of melt could be as thin as 2 km and produce the observed gravity anomaly.

[44] A structure with ~5–20% rhyolite melt at the top of the magmachenber, decreasing with depth to be nearly melt free at a depth of about 15 km is most compatible with our observations. This decrease with depth could be more irregular than the smooth change in the simple velocity model used for the moveout analysis (Figure 4). The positive arrival shortly after the UCN could reflect the base of a fluid-rich part of the magma system, and some of the deeper positive arrivals could be from progressively more melt-poor parts of the crust (Figure 7). A general decrease in melt percentage with depth is most consistent with the amplitudes of the reverberations above the UCN, the estimates of $P$ wave velocities of other studies, attenuation of Moho $Ps$ conversions on rays penetrating the UCN, the gravity anomaly, and the absence of velocity push-down features of the Moho $Ps$ conversion for rays undershooting the UCN.

[45] The mantle reservoir feeding the crustal reservoir is possibly the negative anomaly observed in the upper mantle on cross section 2 near 33 km depth (Figure 7). The arrival is present on the CCP stacks using all rays as well as the stacks made with rays not sampling the UCN. Its presence on the CCP sections not including the UCN rays indicates that it is not a reverberation resulting from the low-velocity

Figure 11. Cartoon showing structure of the Coso geothermal field and the possible relationship between the shallow magma body and regional tectonic features. A new CCP stack along profile 1 combines non-UCN rays using the original velocity structure with UCN rays migrated using the UCN structure of Figure 4. The most important observation is the lack of a lower crustal magma reservoir as determined by the moveout analysis. The magma body may act as a strain guide in the upper crust. This strain guide coupled with lower crustal flow signals a change in deformation mechanism in the Coso area compared to mechanisms believed to exist to the north and possibly the east [Jones and Phinney, 1998].
5.4. Relationship to Regional Tectonics

[46] Previously, Jones and Phinney [1998] inferred the presence of a midcrustal shear zone slightly to the north of Coso based on observations of a midcrustal anisotropic converter that was dipping down beneath the Sierra Nevada to the west. The structure was associated with major shear zones extending down from the large normal faults in the region as drawn by Wernicke [1992], supporting earlier assertions that the area west of Death Valley was deforming under simple shear [Wernicke, 1985; Jones, 1987]. The converter has since been observed farther to the north and east and appears to underlie a large part of eastern California [Jones and Phinney, 1997; Zhou and Phinney, 2000]. Notably, this shear zone underlies a large region lacking major silicic magmatism.

[47] Observations of lithospheric discontinuities in the Coso area provide a unique insight into the interplay of magmatism and tectonics. The anisotropic conversion seen by Jones and Phinney [1998] appears to be absent beneath Coso except for a small region at the north end of the study area. This suggests that this master structure accommodating extension to the north, and the mode of deformation it represents, is not at work within the Coso geothermal area. Either this structure has been hidden by igneous reworking of the crust, or it is absent. Assuming the latter, a different style of crustal extension is required in the Coso region. The deformation regime must satisfy observations that shallow faults seem to root into a decollement on top of the magma chamber (J. R. Unruh, personal communication, 2000) and that the Moho seems nearly planar.

[48] We propose that extension in the uppermost crust occurring on mapped faults is accommodated within the magma body at midcrustal levels (Figure 11). This model requires surface faults to root into the magma body as has been inferred from unpublished active seismic profiles [Unruh et al., 2000, 2001a, 2001b; Pullamanappallil et al., 2001]. Extension beneath the magma is unlikely to be localized directly under the magma body: such extension would raise the Moho many kilometers depending on the amount of extension accommodated in this area. Instead, flow within the lower crust into the region under the magma body could combine with magmatic additions to accommodate extension and preserve the nearly flat Moho observed by this study, much as envisioned by Gans [1987] for the northern Basin and Range. As for the upper crust, widespread extension in the lower crust must root upward into the narrower deforming region of the magma chamber. The broadly deforming lower crust will be separated from the near rigid midcrust to the sides of the magma chamber by a zone with substantial shear. Displacement across this zone would increase toward the magma body. More work needs to be done on the nature of the MCP conversion (anisotropy; density and velocity contrast across the interface), but this conversion seems a possible candidate for such a midcrustal decollement (Figure 11). In particular, its presence near the inferred base of the magma chamber and increase in amplitude approaching the magma chamber seems a good fit to the character expected of such a decollement.

[49] Jones [1987] suggested that the Coso area absorbs differences in tectonic strains between the area of west-northwest extension to the north, left-lateral strike-slip motion to the south along the Garlock fault, right-lateral bending of the Garlock related to the Eastern California shear zone, and the rapidly departing Sierra Nevada block to the west. Such a regime, which is more nearly plane strain than the region to the north and east, might provide a more favorable environment for volcanic processes than surrounding areas. Localization of strain that initiated magmatism could well have reflected the extended step-over geometries of strike-slip faults crossing the region [e.g., Unruh et al., 1998]. Magmatic injections in the Coso area may have originated from a brittle initial response of the crust to the complex tectonic environment. The injections along with lithospheric thinning under the Sierra would serve to heat lower crust throughout the “accommodation terrain,” perhaps making this more pure-shear model of deformation possible.

6. Conclusions

[50] The presence of an upper crustal magma reservoir situated 5 km below the center of the modern Coso geothermal field has been confirmed using receiver function analysis. This reservoir is between 2 and 15 km thick with ≥5% rhyolitic melt. Thinner or more mafic reservoirs require higher melt percentages to satisfy our observations. Receiver function modeling combined with moveout analysis has shown that a lower crustal magma reservoir is unlikely to underlie the Coso geothermal area. A possible candidate for an upper mantle reservoir has been detected near 35 km depth. This mantle reservoir probably feeds the crustal magma body with periodic injections or continuous flow in dikes (< 1 km width). Strain localization in the shallow magma reservoir probably causes the Coso area to extend differently than extensional terranes to the north.

[51] Acknowledgments. This research was conducted with the support of contract N68936-98-C-0234 from the Geothermal Programs Office of the U.S. Navy. Acquisition of the seismograms would have been impossible without the diligent help of many members of that office, most especially Mike Hastings, Rick Webber, and Bob Johnson, and the continued enthusiastic support of Frank Monastero. Brian Zurek, Seth Mueller, Shelly Bolus, Damon Lytle, and Bob Phinney provided needed assistance in the field. Scott Whitehead devised a preamp that greatly improved our data quality. Randy Keller graciously provided a number of seismometers. IRIS PASSCAL provided much equipment on short notice and helped in troubleshooting things in the field. The Ridgecrest Search and Rescue team gleefully drove a snowcat to recover seismometers buried in a late season snowfall. Receiver function modeling would have been considerably more time consuming without the assistance of Anne Sheehan. Discussions with Lang Farmer helped us better understand silicic magma systems, but any errors we retained are in spite of his efforts. Discussions with Jeff Unruh and Bob Phinney improved our analysis and understanding, as have the discussions accompanying the technical symposia run by the Geothermal Programs Office. Jeff Unruh and Jonathan Lees kindly provided material in advance of publication. Reviews by Martha Savage, Jonathan Lees, and an anonymous associate editor clarified, repaired, and tightened our presentation.

References


---

H. J. Gilbert, C. H. Jones, and C. K. Wilson, Department of Geological Sciences, University of Colorado, Campus Box 399, Boulder, CO 80309-0216, USA. (cjones@mantle.colorado.edu; wilsonck@cires.colorado.edu)