The relation of crustal scattering to seismicity in southern California

Justin Revenaugh
Center for the Study of Imaging and Dynamics of the Earth, Earth Sciences Department
University of California, Santa Cruz

Abstract. Dense regional array recordings of 377 teleseismic events are used to map geographic variability in upper crustal P to S scattering in southern California. Scattered-wave energy is sensitive to short-wavelength heterogeneity and maps of scattering potential offer a powerful complement to travel time tomography in characterizing seismic crustal heterogeneity. The scale length of resolved scattering variability is commensurate with the scale lengths of surface fault-trace length and microseismicity variations, and both mapped faults and seismicity are concentrated in regions of strong scattering. Fault-proximal scattering averaged across-strike of predominately strike-slip fault zones is highly correlated with microseismicity levels and is predictive of the pattern of microseismicity variation but not of absolute scale. North-south profiles through the Transverse Ranges reveal coincident strong gradients in microseismicity and scattering, whereas mapped segment bounds along the San Andreas fault zone from San Francisco to the Salton Sea coincide with strong along-strike gradients in the level of scattering. The overall correlation of microseismicity and scattering potential is consistent with structural control of aftershock and background microseismicity production, strain energy control of scattering, or both. Prior evidence for fault offset in the pattern of scattering intensity [Revenaugh and Reasoner, 1997] favors the former.

1. Introduction

Low-frequency (<<1 Hz) wave propagation through the Earth is, on the whole, well approximated by propagation through one-dimensional structures. Attenuation, crustal reverberation, and source signature are the dominant processes affecting waveshape and each is well modeled as a simple convolution operator. Together, they produce body waves in several times the dominant period in duration. The situation is different at high frequencies. Scattering from inhomogeneities within the crust and uppermost mantle produces body wave coda tens to hundreds of times the dominant period in duration. The energy comprising coda follows multiple paths and propagates via multiple modes. While this multiplicity makes deterministic assessment difficult, the resulting space-filling nature of coda energy renders coda decay largely insensitive to variations in source and receiver spacing and details of the first-arrival paths connecting them [e.g., Aki, 1969]. Geographic variation in coda decay rate is documented on scale lengths of 50 to 500 km and appears highly sensitive to local geology [e.g., Phillips et al., 1988] and active tectonics [e.g., Jin et al., 1994]. Temporal variation on subannual [Chouet, 1979; S and Aki, 1990] to decadal timescales [e.g., Sato, 1988; Jin and Aki, 1989] is also noted in conjunction with large earthquakes or variation in seismic b value. In large, coda work has concentrated on local to regional S wave coda recorded at long lag times, that is, long after the passage of the primary S wave.

Studies of early coda cannot rely on the spatial averaging of heterogeneity that late coda provides. Much of this work has centered on crustal reverberation [e.g., Langston, 1989], but there exists a developing body of literature devoted to deterministic mapping of distributed scatterers. The latter can be grouped into four broad categories: (1) F-K analysis of small array data to locate proximal energy sources outside the array (primarily topographic sources of scattered Rg) [e.g., Bannister et al., 1990; Gupta et al., 1990; Wagner and Langston, 1992; Hedlin et al., 1994]; (2) source array studies of local to regional phase coda [e.g., Spudich and Bostwick, 1987; Spudich and Iida, 1993; Matsumoto et al., 1998; Chávez-Pérez and Louie, 1998]; (3) mini-array double beamforming [Rietbrock and Scherbaum, 1999]; and (4) regional array mapping of upper mantle and crustal scatterers [e.g., Nishigami, 1991, 1997, 2000; Revenaugh, 1995a, 1995b, 1995c]. Each approach is data driven, and each makes specific, restrictive assumptions regarding the scattered wave field. The last is potentially the most powerful approach for studying the relation of scattering to seismicity.

With a dense network and many earthquake sources, singly scattered energy within P wave coda can be migrated to its point of origin in the subsurface, allowing the production of short-wavelength, seismic structure maps. Revenaugh [1995a] modified the semblance method of Lay [1987] and Lynnes and Lay [1989] and applied it to a large catalog of teleseismic events to map forward P to P scattering variability in the upper mantle beneath the Southern California Seismograph Network (SCSN), identifying energy scattered from subducted subcrustal lithosphere beneath the Transverse Ranges. P to P scattered-wave imaging is best for targets between 75 and 400 km depth, a window combining strong registration on vertical component instruments and sufficient moveout across the ~4° aper-

Copyright 2000 by the American Geophysical Union.

Paper number 2000JB900304.
0148-0227/00/2000JB900304$09.00
ture of the SCSN, while honoring the predominance of low-angle forward scattering in the earliest P wave coda [e.g., Wagner and Owens, 1993; Wagner, 1997]. P to S scattering is typically wide-angle, favoring shallow targets. Revenaugh [1995b; hereinafter referred to as R95] used teleseismic P to S scattered-wave imaging to map variability of scattering strength in the region surrounding the 1992 Landers, California, earthquake sequence. R95 noted strong correlations between the along-fault variation in scattering strength and the density of aftershock seismicity and, for the main event, inferred coseismic slip at midcrustal depths [Cohee and Beroza, 1994], implying that the structures sensed by scattering exert a strong control on aftershock production. Alternatively, these observations can be taken to imply that preshock strain energy variations affect both the scattered-wave signal and aftershock production or that repeated rupture has produced scattering structures that do not themselves influence seismogenesis but rather act as essentially passive proxies of seismic productivity.

The conclusions of R95 are drawn from inspection of the three fault zones ruptured during the 1992 Landers, California, foreshock-mainshock-aftershock sequence and might be specific to them. In this report I greatly expand the data set and geographic coverage of R95, sampling much of California south of 36°N, including the San Jacinto, Elsinore, and southern San Andreas fault zones and the zone of distributed faulting along the southern flank of the Transverse Ranges. The results corroborate the conclusions of R95 and extend the correlation of seismicity and seismic scattering to southern California in general.

2. Methodology

2.1 Scattered-Wave Imaging

The scattered-wave imaging (SWI hereinafter) technique I use is a variant of the diffraction stacking schemes common to multichannel reflection profiling. In essence, it backprojects scattered energy to its point of origin by distributing it along source-scatterer-receiver isochron surfaces, that is, along the loci of potential scattering points. In theory, energy reradiated from small scattering volumes sums constructively, whereas the sums for seismically transparent loci contain mostly noise and tend to zero with increasing data. In practice, the process is more complex, involving four distinct stages (Figure 1). I discuss these below, beginning with diffraction stacking. Discussion of data preprocessing, the first step, is deferred to a later section.

Due to along-path velocity heterogeneity, frequency- and ray-parameter-dependent propagation effects, variations in instrument and site response, and phase shifts imparted by the scatterer itself, one cannot expect scattered waveforms to sum constructively when backprojected to the scatterer. To surmount the incoherence of scattered waves, SWI operates on running sums of unsigned data, building in tolerance to small perturbations in arrival time and virtual insensitivity to arrival phase. The form the operator takes for the stack value at \( x \), \( S(x) \), is

\[
S^n(x) = \frac{1}{K(x)} \sum_{i=1}^{N} \sum_{j=1}^{M} \left[ \sum_{k=-w}^{w} X_j(x) W(T_{ij}(x,x)) \frac{u_{ij}(t_k - \Delta t)}{2w+1} \right] \frac{1}{n}
\]

Here \( u_{ij} \) is the record of event \( i \) from station \( j \). Time \( t_k \) is referenced to the direct P arrival time \( t_0 \), such that \( t_k = t_0 + k\Delta t \) for digitization rate \( (\Delta t)^{-1} \). Energy from source \( s_j \), scattered at \( x \) and propagated to \( r_j \) as an S wave arrives at time

\[
t_{ij}(x) = t_0 + T_P(s_j,x) + T_S(x,r_j) - T_P(s_j,r_j)
\]

on the seismogram (seismograms are aligned on direct P, such that \( t_0 \) is not a function of source or receiver). Here \( T(a,b) \) is the travel time between points \( a \) and \( b \) of P or S as indicated by

Figure 1. Flowchart of processing and imaging steps of the scattered-wave imaging (SWI) technique. The ultimate output is scattering potential, a unitless statistical estimator of relative scattering intensity.
the subscript. \( W(\tau_{ij}(x,r),x,r) \) and \( X_j(x) \) are windowing operators over time and distance, respectively, that include weighting by geometric spreading and surface obliquity, \( K(x) \) is a normalizing factor equal to the sum of all weights applied to data. Summation over \( l \) is used as a running mean filter on \( u_{ij} \).

We have chosen to use unsigned data rather than the envelope function for numerical efficiency; the two data representations yield essentially identical results. Last, \( n \) specifies the exponent of \( n \)-th root stacking [e.g., McFadden et al., 1986] implemented to enhance the low-energy scattered arrivals.

The windowing operators \( W(\tau_{ij}(x,r),x,r) \) and \( X_j(x) \) are specified as

\[
W(\tau_{ij}(x,t),x,t) = \delta(t - \tau_{ij}(x))H(\tau_{ij}(x) - \tau_c)G(x,r)S(x,r_j)\cos\Theta(x,r_j),
\]

\[
X_j(x) = \begin{cases} 1 & \Delta(x,r_j) \leq \Delta_c \\ -\sin^2 \left( \frac{\pi}{2(\Delta_h - \Delta_c)} \right) & \Delta_c < \Delta(x,r_j) \leq \Delta_h, \end{cases}
\]

where \( H(t) \) is Heaviside’s step function, \( t_c \) is the minimum delay time allowed in the stack, \( G(x,r) \) and \( S(x,r_j) \) are the geometric spreading and surface transmission coefficient of the scattered S phase, \( \Theta(x,r_j) \) is the scattered wave incidence angle at the surface, \( \Delta(x,r) \) is the geocentric angle between the scatterer and station \( j \), and \( \Delta_c \) and \( \Delta_h \) are the corner and cutoff distances of the scattered leg. Data are weighted by, rather than corrected for, \( G(x,r) \) and \( S(x,r_j) \) to reduce the contribution of low signal-to-noise ratio (SNR) data, where SNR is defined as the ratio of singly scattered energy from \( x \) to the sum of other scattered arrivals and ambient noise. Note that (1) is, in essence, a single backprojection step.

Abrupt velocity increases below the scatterer will produce triplications in the travel time curve of \( S \). The Moho is the most obvious velocity jump, but intermittent and variable velocity jumps exist above the Moho [e.g., Heam and Clayton, 1986; Magistrale et al., 1992]. To accurately predict the travel time of \( S \), we must have precise knowledge of these velocity jumps and the smooth variation of velocity between them. Unfortunately, such detailed information is not available for \( S \). In light of this the analysis is limited to forward scattered (albeit wideangle) \( S \) only (hereinafter denoted \( S_g \)) and distances less than 0.6°, by and large avoiding intracrustal head waves and \( S_n \). This has the added benefit of reducing the solid area of scattering sampled at each scatterer.

\( S_g \) times are computed using a Poisson-solid scaled version of the Hadley and Kanamori [1977] \( P \) wave model used in routine location of regional seismicity. Teleseismic \( P \) times are computed using the Hadley and Kanamori crustal model superposed on the IASP91 mantle [Kennett, 1991]. Only the relative move-out of teleseismic \( P \) across the array is needed, and this is only weakly sensitive to our choice of mantle model. Travel time calculation is implemented as a table look-up scheme, greatly reducing the computational burden of stacking, which must evaluate billions of travel times.

2.2 Vertical Resolution

Figure 2a illustrates isochron surfaces of \( P \) to \( S_g \) scattered wave delays from a single scatterer at 10 km depth as illuminated by two widely separated teleseismic sources. The isochrons are stretched along the great circle path away from the source due to finite \( P \) ray parameter (vertically incident \( P \) would produce small circle isochrons centered on the scatterer). This demonstrates how well-distributed sources result in a range of times being sampled for each station-scatterer pair, rendering the stack more robust by minimizing the contribution of crustal reverberations and strong multiple scattering.

Figure 2b illustrates isochrons from two geographically coincident scatterers located at 5 and 15 km depth, respectively, as illuminated by a single earthquake in the Japan subduction complex. Compared to Figure 2a, the travel time differences between the two scatterers are smaller. For shallow \( P \) to \( S_g \) scattering the geographic distance between scatterer and receiver largely dictates delay time; changes in scatterer depth of less than 10 km are important only for proximal stations that, given the station density of the SCSN, comprise only a small fraction of the data contributing to any one stack point.

Figure 2. (a) Isochron surfaces of \( P \) to \( S_g \) scattered wave delay time (from \( P \)) for a single scatterer (open circle) at 10 km depth as illuminated by two widely separated teleseismic sources (insets). Shaded contours are for the event in Tonga-Fiji and are spaced at 5 s intervals. (b) Isochrons from geographically coincident scatterers located at 5 km (solid lines) and 15 km depth (shaded contours), respectively, as illuminated by an earthquake in the Japan subduction complex (inset).
greatly limits the depth resolution of SWI in the $P$ to $S_g$ scattering mode, resulting in maps that are vertical averages of scattering over a depth interval of ~10 km. Depth averaging is not wholly bad, however. First, it reduces a three-dimensional problem to two dimensions if we are interested only in crustal scattering, thereby reducing computation time and model storage requirements. Second, by averaging over the seismogenic zone, statistical comparison of scattering and seismicity along near-vertical faults reduces to essentially one dimension (along-strike distance). In this paper I assume that all scatterers are 10 km below the surface, near middepth of the seismogenic zone, recognizing that the results are depth averages and can be produced by scattering at shallower or deeper depths. Experiments placing the assumed scatterers at the surface and at 20 km depth are also discussed briefly.

2.3 Horizontal Resolution

The horizontal resolution obtainable with SWI is a function of the frequency content of data, station density, and wave velocity. The latter enters both in a mean sense (e.g., scattered $S_g$ has innately higher horizontal resolution than scattered $P_g$) and through lateral heterogeneity that produces phase incoherence across the array. As previously discussed, waveform incoherence is addressed by migrating running sums of unsigned data. The length of the running sum depends on the level of travel time variability and dictates the dimension of the scattering cells employed in migration. To quantify this, assume that the total length of the running sum in (1) is $2w$ seconds, such that scattered energy within $\pm w$ seconds of the predicted time for a scatterer at $x$ recorded at $r_j$ is summed. To isolate horizontal resolution, we will assume further that the depth of scattering is constant. Given that the parent $P$ phase is near vertical in the crust, the running sum includes arrivals originating within a nearly elliptical area geometrically centered on the scatterer. The dimensions of this ellipse vary with separation between the scatterer and receiver but are generally of the order of $w v_{S_g}$ by $6w v_{S_g}$ where $v_{S_g}$ is local $S_g$ velocity and the minor axis is aligned with $(r_j - x)$. Averaging these ellipses over a typical station constellation contributing to $S(x)$ results in a nearly circular scattering cell with radius $r = 3w v_{S_g}$, that is, each estimate of $S(x)$ averages over an area of $9\pi (w v_{S_g})^2$. This averaging sets the lower limit on horizontal resolution. There is no, however, an implicit lower scale length bound on sensitivity: scattering heterogeneities smaller than a scattering cell can be detected although scatterers with dimension close to the illuminating seismic wavelength are most efficient. Heterogeneity at scale lengths much longer than the wavelength of $P$ is essentially transparent, affecting only $P$ wave travel time. For teleseismic data recorded by the SCSN this sets an upper limit of ~20 to 30 km.

2.4 Bootstrapping and Scattering Potential

Since SWI operates on running sums of unsigned data, it is not possible to use waveform coherence to distinguish signal ($P$ to $S_g$ scattered energy) from noise (everything else). It is possible, however, to use energy coherence in the sense of group arrival time. To access information, the probability density function of the stack value for a particular scattering cell in the absence of signal must be determined. This is done by treating $S(x)$ as a regionalized variable [Journal and Huijbenregts, 1978].

The arrival times of scattered energy are given by (2) for the $j$th recording (station $r_j$) of event $i$. At $x$, only a subset of records contributes and the background level is a combination of the noise level on each of these sampled at time $r_j$. Since we cannot isolate the background, we estimate it using the same subset of records and arrival times but ignoring the station index, that is, by shuffling the station affiliations of the contributing records before summing. Randomizing the records has the effect of destroying the coherence of scattered energy while retaining source signatures and the local decay rate of noise. Repeated randomization and summation results in a bootstrap estimate of the distribution of the background level. The true migration sum $S(x)$ is then mapped to significance, that is, the probability that the background level is less than $S(x)$. I refer to this probability as the scattering potential $P(x)$. Scattering potential varies between 0 (lowest significance of scattering) and 1 (highest significance). In the absence of coherent scattered energy, the expected value of $P(x)$ is 0.5, such that half of randomized migration sums are larger and half are smaller. The price paid in this mapping is the loss of absolute scattering strength: maps of scattering potential cannot be used to estimate coda amplitude levels, nor is it possible to directly compare results from region to region. Since I am estimating departure of scattering from the local mean, scattering potential in a region characterized by generally low scattering levels but populated by short-scale variations in scattering strength may appear no different than a strongly scattering area with similar length scales and degree of variation. SWI has a number of advantages over traditional diffraction stacking, however. For example, the wave train of an aftershock in early $P$ coda will not be mistakenly mapped as scattered energy since it moves out slowly relative to $P$ and is included in all the bootstrap sums. Potential is also much less sensitive to the choice of stacking parameters and weighting. The most important advantage, however, is that it can be estimated robustly despite an inability to accurately forward model $S_g$ waveforms.

3. Data

Data used in this study consist of high-frequency, vertical component seismograms of teleseismic $P$ wave coda. Using stations of the SCSN, I obtained over 19,000 records from 377 teleseisms occurring between 1981 and 1997. The distribution of events is shown in Figure 3. Events are chosen on the basis of (1) the number of recordings made by the SCSN, (2) SNR (quantified by the ratio of peak $P$ wave amplitude to mean absolute amplitude in the 10 s window prior to $P$, and (3) source simplicity. Sources deeper than 200 km are preferred since they produce shorter time signatures than shallow events of similar magnitude. In addition, the first arriving depth phase ($pP$) is delayed some 30 s or more relative to $P$, providing a sufficient time window in which to study early $P$ coda without interference. Events shallower than 50 km are also used. For these depths, $pP$ and $sP$ do not move out appreciably relative to $P$ across the SCSN and are close enough in time to $P$ that we can treat the sum of $P$, $pP$, and $sP$ as an extended source signature.

Before stacking, all data are aligned on $P$, decimated to 20 samples per second, and spiked. The latter is accomplished through deconvolution of a generalized source-time function (GSTF) approximated by the linear sum of all records in the event gather. Stacking effectively reduces station-side noise (including crustal scattering) since the array is large with respect to the dominant wavelength of $P$. Little teleseismic waveform evolution occurs over the aperture of the SCSN,
which is small relative to epicentral distance such that takeoff angles vary by less than 2°.

GSTF deconvolution is done with standard wave level techniques and typically produces a P waveform of roughly 0.5 s duration (Figure 4). Since the GSTF is the basic wavelet illuminating scatterers beneath the SCSN, GSTF deconvolution should compress scattered energy as well, leaving waveforms shaped by scatterer response and scattered-leg propagation. These vary from scatterer to scatterer allowing for no further compression of scattered waveforms. For instance, attempts to isolate and deconvolve azimuthally averaged crustal reverberation signatures from the coda of direct P complicates the waveforms of shallowly scattered energy because of variations in ray parameter and propagation mode (Pg, Sg, and Rg).

Crustal reverberations in the early coda of direct P should not significantly bias scattering potential estimates, however, as they do not follow travel time trajectories computed for point scatterers and are averaged out over many stations. As a last processing step, all data is low-pass frequency filtered with a cutoff at 1 Hz and corner frequency of 0.4 Hz. The resulting data have a center frequency of ~0.5 Hz.

4. Validation

Plate 1a shows a synthetic test stack for the area surrounding the 1992 Landers, California, earthquake sequence. To generate this image, synthetic seismograms were computed for a uniform grid of point scatterers and stacked as per (1) through (4) using the parameters in Table 1. The synthetic data set exactly mimics the experimental geometry of the true data set (same source-receiver pairs, same record lengths, same frequency content, etc.) and includes random travel time perturbations of up to 10% of total travel time. Scatterer amplitudes are also subject to random perturbations and singly scattered energy is immersed in appropriately colored random noise with a decay rate equal to the regional mean. Estimates of scattering potential made from fewer than 250 seismograms are zeroed out (white areas in figure).

The synthetic test map gives some indication of SWI horizontal resolution but err toward optimism on several counts. First, although the magnitude of travel time noise in the synthetics is realistic, it is geographically uniform and does not contain the systematic, regional biases that characterize lateral heterogeneity. The latter will be expressed in data images by the mislocation of scattering potential highs and lows. Also missing from the synthetics is multiple scattering, the

<table>
<thead>
<tr>
<th>Table 1. Scattered-Wave Imaging Parameters</th>
</tr>
</thead>
<tbody>
<tr>
<td>Parameter Description</td>
</tr>
<tr>
<td>------------------------</td>
</tr>
<tr>
<td>Cell dimensions, degrees</td>
</tr>
<tr>
<td>Corner distance Δr, degrees</td>
</tr>
<tr>
<td>Maximum distance ΔD, degrees</td>
</tr>
<tr>
<td>Minimum delay time t1, s</td>
</tr>
<tr>
<td>Stacking exponent n</td>
</tr>
<tr>
<td>Running sum duration w, s</td>
</tr>
</tbody>
</table>

Figure 3. Distribution of the 377 teleseismic events (triangles) used in this study. Concentric circles are centered on the SCSN centroid and spaced at intervals of 30° epicentral distance.

Figure 4. (a) Decimated, but otherwise unprocessed, data from the October 7, 1981, m_p 5.9 Fiji Islands earthquake (20.6°S latitude, 178.8°W longitude, 625 km depth). Shown are 8 of 74 traces used in constructing the generalized source-time function (GSTF). (b) The eight traces from Figure 4a following GSTF deconvolution resulting in significant P wave compaction, even for this deep, moderate magnitude event.
most important component of which is low-angle forward scattering. Forward scattered energy arriving both in advance of and in the coda of singly scattered $P$ to $Sg$ results in a lengthened time signature and lowered resolution. A third omission in the synthetic test is systematic variations in scattering intensity as a function of the illumination and scattering angles. $P$ to $Sg$ scattering is not isotropic and even a point scatterer has nodes in its radiation pattern that effectively reduce the number of stations that observe it [e.g., Wu and Aki, 1985]. This will result in some scattering volumes being missed entirely, and essentially all of them being muted, some more than others. The synthetic map is not an attempt to quantify the significance of these effects; rather, it is simply meant to show that the method, given the particular combination of network geometry, the frequency content of data and the mean crustal velocity structure existing in southern California, is theoretically capable of imaging shallow scattering variations at length scales down to 5 km or less.

The map in Plate 1a does not provide an estimate of stack variance or precision. Nor can it be used to synthesize seismograms that could be directly compared to the input synthetic data, making it impossible to compute a meaningful variance reduction. The latter restriction stems from the probabilistic nature of the SWI estimator which does not image absolute scattering strength but instead provides an index of local variability about the regional mean. None of the omitted effects (lateral velocity heterogeneity, multiple scattering, and scattering anisotropy) should result, however, in spurious scattering potential highs or lows. Rather, they should act to broaden highs and lows and drive the image toward the background level, that is, toward $P(x) = 0.5$. The result is a band-pass filtered version of true scattering potential, cut on the high end by the aforementioned effects, and cut on the low end because of the local nature of the estimate. Unfortunately, I have found no way of accurately estimating the properties of the filter, which vary geographically. It is possible, however, to place a bound on the magnitude of scattering potential mislocation due to velocity heterogeneity, which is primarily a function of station coverage and the magnitude of velocity heterogeneity. A conservative upper bound assuming a coherent 10% velocity anomaly and stations covering less than 90° of azimuth is ~4 km, that is, two scattering cells.

The inability to compute variance reduction makes it difficult to validate the method using traditional methods. Plates 1b and 1c document an alternative approach, comparing scattering potential maps for the Landers region computed from nonoverlapping, nonequal size portions of the data set partitioned by event date. By chance the number of teleseisms in the data set occurring on odd-numbered days of the month greatly exceeds the number of events on even-numbered days, producing a roughly 60%-40% split of the data set. As can be seen from Plate 1, the location, magnitude and shape of scattering potential highs and lows are consistent between the two images. Although there is a tendency for the odd-days image to be more saturated (scattering potential values closer to zero and unity), reflecting the greater data density and lower bootstrap variance, there is no evidence that the results are dominated by unusual events or noise bursts. Images obtained by removing subsets of stations from the stack produce similar results, confirming that the image is not dictated by a few noisy stations. Last, Figure 5 shows data arrayed as a function of $P$ to $Sg$ delay time from the scattering cell indicated in Plate 1, filtered as per (1) and stacked into small delay-time bins. A high-pass filter has been applied to remove the slow exponential decay of mean coda level. A clear energy pulse is seen to move out along the predicted trajectory, confirming the existence of significant energy re-radiated from a strong scatterer. While these tests fall short of the rigorous validation possible for tomographic images, they do afford confidence that the images produced by SWI are stable, repeatable, and the product of coherent energy in the data.

5. Results

Scattering potential was estimated for the portion of southern California extending from 115° to 121°W and from 32.5° to 36°N. Only cells with 250 or more contributing seismograms were retained, and synthetic tests, such as Plate 1a, were used to eliminate remaining cells with poor scattering potential resolution. Together, these criteria reduced coverage by roughly 45%. In the following, I focus on three subsections: (1) the area surrounding the 1992 Landers earthquake sequence in the eastern California shear zone, (2) a region encompassing the Elsinore and San Jacinto fault zones and the southernmost segment of the San Andreas in California, and (3) a block including the Western and Central Transverse Ranges that extends south of the Palos Verdes peninsula. The $P$ to $Sg$ scattering potentials for each area are shown in Plates 2, 3 and 4, respectively. As discussed above, these maps assume a uniform scatterer depth of 10 km, but because of the vertical averaging inherent to scattered-wave imaging, potential at a point is a weighted sum of scattering occurring between ~5 and ~15 km.

Evident to those familiar with southern California basement geology is the lack of clear expression of geological contacts in scattering potential. Also apparent is the lack of strong visual correlation of scattering with the location of mapped
Plate 1. (a) Test image for the area surrounding the 1992 Landers earthquake sequence computed using parameters in Table 1 and synthetic data as described in the text. Asterisks mark positions of isolated scatterers at 10 km depth. Scattering potential is color contoured and ranges from 0 (lowest likelihood of scattering) to 1 (highest likelihood). Blank areas have fewer than 250 contributing seismograms. Most scatterers are detected and well isolated from their neighbors. (b) Split data test using events from even-numbered days of the month. Open circle marks position of a strong scattering examined in Figure 5. (c) Split data set from odd-day events. The data-derived images in Plates 1b and 1c share no data but are very well correlated, documenting a stable estimator.
Plate 2. Scattering potential for region surrounding the 1992 Landers earthquake sequence with magnitude $M_L \geq 2$ seismicity (black triangles) occurring between 1981 and 1997 (inclusive) superimposed. Symbol size scales with the logarithm of the number of earthquakes. Bold green and black lines represent fault zones used in across-strike averaging discussed in text; colored bars are 10 km long. White areas have fewer than 250 contributing seismograms or poor resolution in synthetic tests.
Plate 3. (a) Synthetic test for the region surrounding the San Jacinto and Elsinore fault zones. All other details of the image as in Plate 1a. (b) Scattering potential with magnitude $M > 2$ seismicity occurring between 1981 and 1997 (inclusive) superimposed. All other details as in Plate 2.
Plate 4. (a) Synthetic test for the region surrounding the western and central Transverse Ranges. All other details of the image as in Plate 1a. (b) Scattering potential with magnitude $M_I \geq 2$ seismicity occurring between 1981 and 1997 (inclusive) superimposed. All other details as in Plate 2.
surface fault traces. Although the precise position of faults at depth cannot be inferred from their surface traces, there is little evidence of consistently high or low scattering along major faults and some scattering potential patterns cross faults with little apparent change in shape or magnitude. These observations raise the question of what precisely is being imaged by SWI. It is clear that SWI as implemented here images only a subset of shallow scattering structure. In part, the search for correlation between scattering and seismicity and other independent observables that makes up most of this report is an attempt to ascertain just which subset.

5.1 Regional Variation

Although the primary purpose of this paper is to document correlation between scattering potential and microseismicity measured along known faults, scattering potential is estimated off-fault as well as on-fault. Here I discuss properties of the scattering field as a whole. In this context the most basic question one can ask is this: Does scattering potential vary at length scales similar to topography, faulting, and seismicity, and, if so, does it covary with any of them? The answer is yes for faulting and seismicity, with both concentrated in regions of high scattering potential.

To quantify these relations, the correlation length scales of scattering potential, topography, microseismicity and surface fault trace length were estimated by the semivariogram offset at 80% of total variance [e.g., Journel and Huijbregts, 1978]. The resulting length scale of scattering is 8 km. Prior to computing the correlation length scales of microseismicity and faulting, both fields were resolved on the 0.02° by 0.02° grid used for scattering, where the number of $M_L \geq 2$ events from 1981 through 1997 and the length of mapped fault traces falling within a grid cell, respectively, were posted at the grid center (the fault map used is shown in Plates 2 through 4 [Jennings, 1975]). The offset at 80% of semivariogram total variance (or 80% of the " sill ") is ~8 km for seismicity and ~5 km for faulting, both very close to the scale length of scattering, and both much less than the ~28 km scale length of topography computed from a 1-minute digital elevation model [Revenaugh, 1995a]. This indicates that resolved shallow P to $S_g$ scattering potential varies at length scales comparable to seismicity and faulting and warrants further analysis.

The linear correlation coefficient between scattering potential and surface fault-trace length is 0.023 and nearly identical to that between scattering and $M_L \geq 2$ seismicity (0.022). Such low values of correlation coefficient are to be expected. Only about one in nine of all scattering grid cells has posted $M_L \geq 2$ seismicity and fewer have mapped faults; any variation of scattering in the remaining cells, indicative of off-fault heterogeneity, must reduce the correlation. Spearman’s Rank correlation method yields larger coefficients (Table 2), revealing a tendency of faults and microseismicity to cluster in regions of high scattering potential. To better investigate this tendency, I have investigated the cumulative density functions of faulting and seismicity as a function of scattering potential.

<table>
<thead>
<tr>
<th>Observables</th>
<th>Statistic</th>
<th>Correlation Coefficient Significance, or Maximum Deviation %</th>
</tr>
</thead>
<tbody>
<tr>
<td>P(x), faulting</td>
<td>Linear correlation</td>
<td>0.023 &lt;1</td>
</tr>
<tr>
<td></td>
<td>Rank correlation</td>
<td>0.410 1</td>
</tr>
<tr>
<td></td>
<td>Kolmogorov-Smirnov</td>
<td>0.048 3</td>
</tr>
<tr>
<td>P(x), $M_L \geq 2$</td>
<td>Linear correlation</td>
<td>0.035 &lt;1</td>
</tr>
<tr>
<td></td>
<td>Rank correlation</td>
<td>0.185 &lt;1</td>
</tr>
<tr>
<td></td>
<td>Kolmogorov-Smirnov</td>
<td>0.111 &lt;1</td>
</tr>
<tr>
<td>P(x), $M_L \geq 3$</td>
<td>Linear correlation</td>
<td>0.024 2</td>
</tr>
<tr>
<td></td>
<td>Rank correlation</td>
<td>0.439 2</td>
</tr>
<tr>
<td></td>
<td>Kolmogorov-Smirnov</td>
<td>0.101 3</td>
</tr>
<tr>
<td>P(x), $M_L \geq 4$</td>
<td>Linear correlation</td>
<td>0.015 5</td>
</tr>
<tr>
<td></td>
<td>Rank correlation</td>
<td>0.486 17</td>
</tr>
<tr>
<td></td>
<td>Kolmogorov-Smirnov</td>
<td>0.106 10</td>
</tr>
</tbody>
</table>

Figure 6. Fraction of total surface fault-trace length occurring in scattering cells with $P(x) \leq s$ (solid line) versus the fraction of scattering cells with $P(x) \leq s$ (dashed line). If surface faulting and scattering are independent processes, the two curves should overlap with small differences due to finite sample size. Surface fault traces, however, concentrate in regions of strong scattering, resulting in a systematic mismatch of the two curves.
levels as high as $M_f = 3$ (Figure 9). The increase in seismicity with scattering potential is less striking for $M_f \geq 4$ seismicity and absent for $M_f \geq 5$. Whether this reflects a true change in the spatial relationship of scattering and seismicity of this magnitude or simply a temporally too-short catalogue of seismicity is unclear. What is clear is that high scattering potential cells are more likely to have microseismic activity than low scattering potential cells.

5.2 Along-Strike Variation

The preceding analysis exploited the vertical averaging properties of $P$ to $S_p$ migration by ignoring the depth of seismicity (all events were shallower than 22 km). Horizontal averaging entered only in the binning of seismicity, which was at a level commensurate with the standard error of event location. Here I follow R95 by restricting attention to only fault-proximal scattering and employing across-strike horizontal averaging. As will be shown, the exclusion of off-fault variability greatly increases the predictive power of the correlations.

Figure 10 illustrates the geometry used for across-strike averaging. In all subsequent examples, $P(x)$ is averaged and microseismicity counted in boxes measuring 4 km along fault strike and 15 km across strike, such that a typical box contains 30 scattering cells. Boxes are spaced at 2 km intervals along the fault. Fault curvature causes the boxes to overlap off the fault. Where this occurs, scattering cells are averaged and microseismicity counted for all boxes they fall within.

Across-strike averages for the Joshua Tree, Landers and Big Bear aftershock zones are shown in Figure 11 for the fault zones indicated in Plate 2. Similar estimates were presented in R95. Refinements in SW1 methodology and a nearly threefold increase in data produce some changes, most notably with re-

Figure 7. (a) Relative likelihood that a scattering cell is crossed by a surface fault trace [Jennings, 1975], showing the increasing likelihood of faulting with increasing scattering potential. (b) Relative length of surface fault traces in scattering cells as a function of scattering potential showing a monotonic increase with increasing scattering potential.

sis that scattering and surface faulting are independent at the 3% significance level.

Figure 7 provides additional insight into the relationship between surface faulting and scattering. Here both the likelihood of faulting and the length of faults in scattering cells is shown relative to the mean over six scattering potential ranges (0.0 to 0.1, 0.1 to 0.3, 0.3 to 0.5, 0.5 to 0.7, 0.7 to 0.9, and 0.9 to 1.0). Both faulting statistics increase monotonically with increasing scattering potential, accumulating a roughly 25% increase in the likelihood of faulting and length of fault trace over the range of scattering potential. In other words, surface fault traces are roughly 25% more likely to occupy high scattering potential cells than low scattering potential cells.

Similar relationships hold for seismicity. Although the enrichment of seismicity in high scattering potential cells is most statistically significant for the lowest magnitude threshold ($M_f \geq 1$) (Figure 8 and Table 2), the roughly 30% increase in the likelihood of seismicity and cumulative seismicity over the range of scattering potential holds for minimum magnitude

Figure 8. Fraction of total $M_f \geq 1$ seismicity between 1981 and 1997 (inclusive) occurring in scattering cells with $P(x) \leq s$ (solid line) versus the fraction of scattering cells with $P(x) \leq s$ (dashed line). Microseismicity concentrates in regions of strong scattering, resulting in a systematic mismatch of the two curves.
Figure 9. (a) Relative likelihood that a scattering cell was seismically active between 1981 and 1997 for magnitude thresholds between $M_L \geq 1$ and $M_L \geq 3$, showing the increasing likelihood of seismicity with increasing scattering potential. (b) Total seismicity normalized by mean seismicity as a function of scattering potential. The number of earthquakes exceeding each magnitude threshold generally increases with increasing scattering potential. No trend is apparent for $M_L \geq 4$ events (not shown).

spect to the Big Bear aftershock zone, where the correlation of mean scattering potential and microseismicity is now positive (see Table 3). The negative correlation of scattering potential and mainshock slip [Cohee and Beroza, 1994] noted in R95 remains (not shown). I did not examine aftershocks of the October 16, 1999, Hector Mine event due to the still-evolving aftershock catalog and poor scattering potential resolution in the vicinity of the mainshock.

Across-strike averages for the San Jacinto and Elsinore fault zones are shown in Figure 12. Along these nearly 200 km faults zones the seismicity statistics are dominated by distinct clusters of aftershocks and background seismicity. Because seismic productivity varies from segment to segment and because the completeness of the catalog varies with station coverage, I have chosen to use the logarithm of cumulative microseismicity. This has the effect of reducing multiplicative bias to additive bias, that is, of reducing along-strike changes in scale between potential and microseismicity to baseline offsets. It also better honors the regional nature of $P(s)$ which does not contain wavelengths longer than ~60 km. Although the correlations are poorer than the 1992 Landers sequence events, they are highly significant. Poorer performance for these two fault zones may be due to the incompleteness of the

Table 3. Across-Strike Averaged Profile Correlations

<table>
<thead>
<tr>
<th>Fault Zone</th>
<th>Length (km)</th>
<th>Linear Correlation</th>
<th>Significance, %</th>
<th>Rank Correlation</th>
<th>Significance, %</th>
</tr>
</thead>
<tbody>
<tr>
<td>Landers</td>
<td>80</td>
<td>0.68</td>
<td>&lt;1</td>
<td>0.64</td>
<td>&lt;1</td>
</tr>
<tr>
<td>Joshua Tree</td>
<td>38</td>
<td>0.52</td>
<td>7</td>
<td>0.49</td>
<td>7</td>
</tr>
<tr>
<td>Big Bear</td>
<td>32</td>
<td>0.64</td>
<td>1</td>
<td>0.67</td>
<td>&lt;1</td>
</tr>
<tr>
<td>San Jacinto</td>
<td>180</td>
<td>0.32a</td>
<td>4</td>
<td>0.23</td>
<td>9</td>
</tr>
<tr>
<td>Elsinoreb</td>
<td>165</td>
<td>0.46a</td>
<td>&lt;1</td>
<td>0.45</td>
<td>&lt;1</td>
</tr>
</tbody>
</table>

a Log seismicity.

b $M_L \geq 1$ seismicity.
Figure 11. Across-strike average scattering potential (thin line) and $M_L \geq 2$ seismicity (bold line) profiles for the (a) Joshua Tree, (b) Landers, and (c) Big Bear events of the 1992 Landers earthquake sequence. Averaging is performed along the thick green and black fault lines shown in Plate 2. For all three profiles, the origin is fixed at the southernmost end. See Table 3 for details of the correlations.

Figure 12. (a) Across-strike average scattering potential (thin line) and $M_L \geq 2$ seismicity (bold line) profiles for the San Jacinto fault zone. (b) Across-strike average scattering potential (thin line) and $M_L \geq 1$ seismicity profiles (bold line) for the Elsinore fault zone. A lower magnitude threshold is chosen due to the low cataloged seismicity on the Elsinore fault. Averaging is performed along the thick green and black lines shown in Plate 3. For both profiles the origin is fixed at the southernmost end. See Table 3 for details of the correlations.
mainshock catalog along each fault, resulting in gaps or lows in seismicity uncorrelated with scattering potential. The Anza gap is particularly noteworthy in this regard, however, as it is marked by a low in seismicity and the most pronounced scattering low observed in the study area. I return to this point later.

The linear and rank correlations of mean potential and cumulative microseismicity (or log seismicity) are positive for each fault zone. Using Fourier phase randomization, the likelihood of equal or greater chance linear correlation was evaluated for each and found to range from ≤ 1% for the Landers, Big Bear and Elsinore fault zones, to 7% for Joshua Tree (Table 3). For reasons mentioned above, linear correlation may not be the best statistic, but the fact that it is positive and significantly nonzero for all of the primarily strike-slip faults tested is diagnostic of (1) structural control of microseismicity production, (2) sensitivity of both scattering and microseismicity to strain energy, perhaps through the opening and closing of cracks, or (3) production of "passive" scattering structures by repeated rupture. In any case, across-strike averaged scattering potential is predictive of the pattern of microseismicity variation, but not the absolute scale. This is true of fault segments ruptured in a single event (e.g., Joshua Tree) and for long faults comprising many segments (e.g., San Jacinto fault zone).

5.3 Effect of Scatterer Depth

Experiments placing the assumed scatterers at the surface and at 20 km depth were performed and compared with seismicity as described previously. In both experiments the level of correlation was reduced for the five strike-slip-dominant fault zones examined in this study, although the positive sense of correlation remained. For instance the linear correlation of across-strike averaged $M_L \geq 2$ seismicity and scattering along the Landers fault zone (see Plate 2) drops from 0.68 to 0.29 for scatterers at the surface and to 0.56 for scatterers at 20 km depth. Along the Big Bear fault zone, linear correlation drops from 0.64 to 0.15 and 0.29 for scatterers at the surface and 20 km depth, respectively, and similar magnitude drops in correlation are observed for the other three fault zones in Table 3. Although depth resolution is limited, these results nonetheless suggest that scattering within the seismogenic zone (~5 to ~15 km depth) correlates best with seismicity.

5.4 San Andreas Fault

The San Andreas fault is marked by very low levels of background seismicity throughout the study area. The exception is seismicity within the San Gorgonio pass, in particular, seismicity related to the 1986 North Palm Springs earthquake. To the north, large gaps in seismicity exist that make meaningful correlation with scattering potential impossible unless one accepts that the catalog record is indicative of the long-term average seismicity along the fault, a difficult assumption to accept given the short earthquake catalog and the long recurrence intervals of some San Andreas segments [Sieh et al., 1989]. It is possible, however, to investigate the relation between scattering potential and fault segmentation [e.g., Revenaugh and Reasoner, 1997]. In Figure 13, the segment bounds of the Working Group on California Earthquake Probabilities (WGCEP) [1988], Jones [1988], and Wyss and Zhong [1995] are plotted against scattering potential along a portion of the San Andreas fault. Coverage is extended north of the present study area using SWI images for central California from Revenaugh and Reasoner [1997]. Scattering is seen to vary greatly over the ~800 km of imaged fault, demonstrating laterally extensive highs and lows. Of special note is the correspondence of high along-strike gradients in scattering and segment bounds and the fact that only four segment bounds fall in low-gradient regions. High along-strike scattering gradients are numerous, however, invoking the specter of coincidence. There is also the question of segment bound mislocation.

To assess the significance of association of segment bounds with high along-strike gradient, I first estimated the probable error associated with segment bound location. Based on fault bends, initiation and termination of historic large earthquake and along-fault focal mechanism or stress variability, segment bounds are likely accurate only to ±10 km. This number corresponds to the median separation of segment bounds contained in both the WGCEP and Jones [1988] catalogs and is used in evaluating the proximity of a segment bound to high scattering gradient by conducting the search within a 20 km window centered on the published segment location.

High along-strike gradient is defined by a threshold which I varied from 0.03 km$^{-1}$ to 0.07 km$^{-1}$ (note that scattering potential is unitless). Figure 13a assumes a threshold of 0.05 km$^{-1}$, approximately the standard deviation of along-strike gradient magnitude. At this level, 13 of the 17 distinct segment bounds lie within ±10 km of high scattering gradient, whereas only 52.4% of the fault zone as a whole lies this close to high gradients (Figure 14). The probability of observing 13 or more chance associations is less than 4%. Although significance varies with gradient threshold and with choice of segments, for the range of combinations examined only one set fails to be significant at the 10% level (Figure 14).

This result corroborates that of Revenaugh and Reasoner [1997] for central California and bolsters the conclusion that the correspondence of segment bounds and scattering gradients is not coincidence, that is, that most segment bounds have seismically mappable structural expression. Whether this structure is generative or derivative cannot be determined from scattering alone. It is noteworthy, however, that segment bounds fall within scattering gradient highs, not on scattering peaks or troughs. This suggests that the aspects of segment bound structure imaged in this study do not straddle the boundary but rather lie to one side or the other along strike.

Where seismicity is most abundant along the southern San Andreas fault, namely, in the San Bernardino-San Gorgonio region, stress segmentation on the basis of focal mechanism variability is most refined [Jones, 1988; Wyss and Zhong, 1995; Seebier and Armbruster, 1995; Magistrale and Sanders, 1996]. This stretch of the San Andreas has a high density of segment bounds and few unpaired high scattering gradients. Along the Mojave and Coachella Valley stretches, seismicity is less frequent, segments are longer, and a number of unpaired high along-fault scattering gradients are found (Figure 13a). While it is attractive to think that additional segment bounds exist and that scattering can be used to identify them, such is not the case. While it appears that most mapped segment bounds are expressed in scattering, not all scattering gradient highs are segment bounds, as evidenced by the abundant off-fault variation of scattering potential. Which along-strike highs in scattering gradient correspond to bounds and which are unrelated structure cannot be determined from scattering alone.
Figure 13. (a) Across-fault averaged scattering potential profile for the fault zone shown in Figure 13b. Origin is fixed at the northernmost end. Included are positions of segments bounds of the WGCEP, Jones [1988] (marked J), and Wyss and Zhong [1995] (marked W where different from Jones [1988]). Segment bounds are dashed if they fail to lie within 10 km of high along-fault scattering gradients (defined here as any gradient magnitude ≥ 0.05 km¹). Of the 17 distinct boundaries, 13 lie on or very near high-gradient patches. Near-coincident WGCEP and Jones segment boundaries near 500 km and 630 km are grouped and placed at the geographic mean position. NC, north coast; SCM, Santa Cruz Mountains; PF, Parkfield; SGP, San Gorgonio Pass; CV, Coachella Valley. (b) Fault zone used in Figure 13a. Shaded bars are 50 km long.

Figure 14. (a) For each threshold gradient the number of segment bounds $h_0$ within 10 km of high along-fault scattering gradient is shown for the WGCEP bounds (solid circles) or the combined WGCEP, Jones [1988], and Wyss and Zhong [1995] bound sets (open circles). Gray bars mark the total number of distinct segments (13 and 17, respectively). The percentage of total fault length within 10 km of high along-fault scattering gradient (solid line) is shown on the right-hand axis and decreases with increasing threshold gradient. (b) The probability of chance observation of $h_0$ segment bounds within 10 km of high along-fault scattering gradient is shown for the segment bound sets in Figure 14a. In only one case is the observed correspondence more than 10% likely for randomly placed segments.
5.5 Transverse Ranges

Plate 4 reveals a near-continuous high scattering potential band along the southern flank of the western and central Transverse Ranges. Seismicity is largely confined to this band, most densely populating the southernmost limits of high scattering.

Across-strike averaging of scattering potential and seismicity in this area is made difficult by extensive thrust faulting, for which both along-dip and along-strike variability is important. To better honor this, I have chosen to subdivide the region into 10 north-south trending corridors measuring 20 by 65 km (Plate 4). Scattering potential is averaged and seismicity is counted in bins spanning the 20 km width of the corridor and extending 4 km along corridor. The results, shown in Figure 15, reveal a clear correspondence between seismicity and scattering. In particular, seismicity peaks in regions of strong scattering. There is an apparent asymmetry to the relationship: while the southern onsets of high scattering and seismicity are generally coincident, scattering remains at high levels 20 km or more north of the peak in seismicity. Grouping the results into a single scattergram reveals the general correspondence of scattering and seismicity peaks (Figure 16).

For the corridors shown in Figures 15e, 15f, and 15g, the onset of peak seismicity and scattering marks the southern topographic front of the Transverse Ranges. This could indicate a strong topographic component of scattering, rendering the apparent correlation of scattering and seismicity a simple coincidence of geography and perhaps explaining the northward extension of high scattering (beyond the high in seismicity) into the topographically rugged ranges. Evidence of topographically scattered $R_g$ waves is presented by Revuenaugh [1995c] and Revuenaugh and Mendoza [1996] for southern California, and we must assume that at least some component of scattered $S_g$ is created by topography. However, closer examination of Figure 15 reveals that the topographic front of the Transverse Ranges does not correspond with peak seismicity or scattering for the remaining corridors. While it is surface roughness, not elevation, that produces scattered energy, elevation is an excellent proxy for roughness in the area of interest. Last; the apparent correlation is reduced when the assumed depth of scattering is \leq 1 km, that is, where topographic scattering occurs. Thus it appears unlikely that topographic scattering is dominant in the images; rather, it is subsurface structure that produces scattering which, in pattern, corresponds to seismicity.

![Figure 15](image)

**Figure 15.** Mean scattering potential (thin line) and total $M_L \geq 2$ seismicity (bold line) for 10 north-south profiles crossing the Transverse Ranges. Profiles are shown Plate 4 and are arrayed west (Figure 15a) to east (Figure 15j). For most, the location of greatest south-to-north increase of scattering potential and seismicity coincide.
This study documents a new line of evidence relating fault zone structure to seismogenesis. Although I have not uniquely related seismic scattering to specific aspects of subsurface structure, it is by definition a product of elastic heterogeneity and by demonstration a pattern predictor of microseismic productivity. In the latter sense, near-fault scattering potential and velocities are quite similar. When attention is restricted to known active fault zones, each displays a correlation with seismicity that lacks an implicit scale: a particular velocity anomaly is not associated with a prescribed level of aftershock seismicity or mainshock slip, nor does scattering scale uniformly with seismicity throughout southern California. Despite this similarity, the relation of the two predictors, velocity and scattering, to each other is somewhat problematic. High seismic productivity is correlated to both high subsurface velocities and high scattering, suggesting that high-velocity material is intrinsically strongly scattering. Given our ignorance of the source of scattering, it is difficult to rule this out, but there is no obvious a priori reason to suspect that scattering intensity should scale with velocity. Thus it is not clear that tomography and SWI predictions are founded on the same aspects of subsurface structure.

My results support previous claims that segment boundaries are expressed in scattering and thus associated with distinct subsurface structures [Revenaugh and Reasner, 1997; Nishigami, 2000]. The nature of scattered-wave imaging leaves the precise form of this expression unresolved, but one can conjecture as to its cause. Many segment bounds are associated with fault zone bends and gaps. Complex faulting and fracturing around these junctures [e.g., King, 1986] might presumably find expression in scattering, such that scattering images a secondary, or produced, structure. Alternatively, scattering may image causative structures that predate faulting, or a mix of both. A similar "chicken and egg" problem is raised by the pattern predictive relation of scattering to microseismicity: Does the imaged structure modulate seismogenesis or is it the product of repeated rupture? A third possibility is that scattering structure is unrelated to seismogenesis, but both it and microseismicity are proxies of stress energy. Such might be the case for a field of distributed cracks which open and close in response to regional stress variations.

Revenaugh and Reasoner [1997] computed cross-fault averages of scattering potential separately for each side of the San Andreas fault in central California. Following modest filtering to remove poorly constrained long-wavelength variability, these profiles were compared after the removal of variable amounts of right- and left-lateral offset. Optimal correlation was achieved with the removal of 315 km of right-lateral offset. This value agrees well with geologic estimates of cumulative offset [e.g., Matthews, 1976], suggesting that the greater portion of scattering is essentially frozen into the crust and likely predates all but the initial stages of faulting in early Miocene time. In particular, it seems especially unlikely that correlated scattering patterns separated by 315 km would be due to present-day stress field variations. For the San Andreas at least, it appears that the scattering structure is very old. The San Andreas, however, displays no correlation between scattering and seismicity, unlike the other five fault zones examined herein. Whether or not scattering is largely static along them is uncertain but an important question.

As mentioned previously, the Anza stretch of the San Jacinto fault zone is marked by a historic gap in seismicity [Thatcher et al., 1975] and very low scattering potential. If
scattering is frozen into the crust or the product of repeated rupture, the general correlation of scattering and microseismicity along the San Jacinto suggests that time-averaged seismic productivity at Anza is low and that any future gap-filling event will not greatly elevate long-term productivity. If, on the other hand, scattering responds to the regional stress field, the present low scattering potential observed at Anza suggests continued quiescence. This conjecture follows from R95 which demonstrated strong correlation of seismicity following the 1992 Landers earthquake sequence and scattering potential prior to the mainshock. For that sequence at least, scattering in regions of high aftershock density was elevated at least 10 years prior to rupture.

Acknowledgments. This work was supported by NSF award EAR-9614874. The author thanks the Southern California Earthquake Center whose remarkable data archives make possible this study. Anton Dainty, Michael Hedlin, and Scott Phillips provided thorough, thoughtful, and very helpful reviews.

References

Nishigami, K., Spatial distribution of coda scatterers in the crust around two active volcanoes and one active fault system in central Japan: Inversion analysis of coda envelope, Phys. Earth Planet. Inter., 104, 75–89, 1997.


J. Revenaugh, Center for the Study of Imaging and Dynamics of the Earth, Earth Sciences Department, University of California, Santa Cruz, CA 95064. (jsr@es.ucsc.edu)

(Received October 25, 1999; revised July 10, 2000; accepted August 4, 2000.)