GEOLOGIC APPLICATIONS OF SEISMIC SCATTERING

Justin Revenaugh
Earth Sciences, University of California, Santa Cruz, CA 95064; jsr@monk.ucsc.edu

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ABSTRACT
Once disregarded as noise, scattered seismic waves are finding increasing application in subsurface imaging. This sea change is driven by the increasing density and quality of seismic recordings and advances in waveform modeling which, together, are allowing seismologists to exploit their unique properties. In addition to extensive application in the energy exploration industry, seismic scattering is now used to characterize heterogeneity in the lower continental crust and subcrustal lithosphere, to examine the relationship between crustal structure and seismogenesis, and to probe the plumbing of active volcanoes. In each application, the study of seismic scattering brings wavelength-scale structure into sharper focus and characterizes the short scale-length fabric of geology.

INTRODUCTION
The notion of a horizontally stratified, or layer-cake, Earth pervaded seismology for many years, its prevalence as much a reflection of seismologists’ unwillingness to surrender analytical and numerical simplicity as of gravity’s dominance in geologic processes. With time, the intersection of need and increasing computing power produced successive generalizations of seismic structure, each seeking to explain a greater percentage of the recorded wave field and to extract more information from the same wiggles. The products of the latest stage of this evolution—three-dimensional reflection profiling and tomography—are spectacular and very much a part of geology.

Only a few holdovers of the layer-cake paradigm remain. The most important is scale-length separation. The subsurface is treated not as a true three-dimensional continuum but rather as a collection of short-wavelength, laterally
continuous features (reflectors) superposed on a smoothly varying background, splitting the heterogeneity spectrum in two and forcing different, but quite limited, forms of regularity on each half. The efficacy of this approach in everyday practice is sufficient validation. Nonetheless, the evidence of its failure appears on every seismogram, manifest, for instance, as the slowly decaying coda trailing “primary” arrivals in time. Created by a multiple-scale, laterally varying seismic structure that does not lend itself to the former parameterization, the scattered arrivals composing coda are unfairly treated as “noise,” that is, as an unavoidable nuisance. That attitude began to give way in the 1970s with the pioneering work of K. Aki (e.g. Aki 1969, Aki & Chouet 1975). Recognizing coda of regional S waves as the sum of diffusely scattered body and surface waves, Aki’s work (Aki 1969) introduced an alternative parameterization of Earth structure, one based on the statistics of velocity and density fluctuations rather than on deterministic description (see also Chernov 1960).

Statistical characterization of the subsurface is extremely powerful, describing aspects of the subsurface that, if usually not dominant, are at least always present. For many wave propagation regimes (e.g. turbulent media) it is the only natural parameterization. Information about this portion of the heterogeneity spectrum comes almost entirely from scattered waves, that is, from the waves produced by primary wave interaction with it. This review focuses on recent geologic applications of seismic scattering with specific emphasis on the structural fabric of the lower continental crust and subcrustal lithosphere, regions of active volcanism, and the relationship of crustal heterogeneity to the seismogenic process.

THEORETICAL BACKGROUND

Throughout much of the Earth, short scale-length fluctuations of velocity and density about the mean are small, typically a few percent or less. For such a weakly inhomogeneous medium, it is possible to separate the primary wave field (waves that would exist in the medium without any fluctuations) from the scattered wave field (waves generated by primary wave interaction with fluctuations). The ability to separate wave field components is largely responsible for the success of the deterministic parameterization which explains only the primary wave field.

Properties of the scattered wave field depend on both the primary wave field and the heterogeneity. Scattering is most efficient when the wavelength of primary waves, \( \lambda = \frac{2\pi}{k} \), where \( k \) is wavenumber, is commensurate with the scale length of heterogeneity, \( a \), that is, when \( ka \approx 1 \). For \( ka \gg 1 \), velocity varies little over a wavelength, and the medium appears essentially homogeneous. The travel time and amplitude of primary arrivals vary, but scattered waves carry little energy. Conversely, for \( ka \ll 1 \), velocity varies so frequently
over a wavelength that wave energy is held entirely by the scattered wave field, in which case long-range propagation is modeled accurately by primary wave propagation in an equivalent homogeneous medium. Between these two extremes, the scattered wave field can be very complex.

It is common to imagine scattering as wave field interaction with a distribution of discrete scatterers, allowing a separation of the scattered wave field into singly and multiply scattered components, distinguishing between scattering of the primary and scattered wave fields. For weakly inhomogeneous media, singly scattered energy dominates, and multiple scattering becomes important only at long lapse times (arrival times much longer than that of the primary waves). Singly scattered waves are further classified as either forward scattered or back scattered. Forward-scattered waves travel subparallel to the primary waves and are the dominant component of early $P$-wave coda. Back-scattered waves propagate at high angles to the primary wave field and account for a larger percentage of $S$-wave coda. Phase conversion (e.g. $P$ to $S$ wave and body to surface wave) is an important process, producing a rich sequence of secondary arrivals sensitive to compressional and shear velocity, density, and roughness of internal boundaries and the free surface.

Figure 1 shows a typical regional array recording of a teleseism in which a slowly decaying coda continues long after passage of the primary $P$ wave. Variation from station to station proves that this late-arriving energy is generated locally in the lithosphere beneath the stations, but even the best three-dimensional deterministic models predict little of this energy. Missing from those models are scale lengths of heterogeneity comparable to seismic wavelength: scale lengths that are averaged over or healed out of primary waves propagating many wavelengths (Gudmundsson 1996, Snieder & Lomax 1996) but that are the most effective scatterers.

CONTINENTAL LITHOSPHERE

A number of authors have attempted to relate teleseismic $P$-wave coda levels, as well as amplitude and phase fluctuations of the primary $P$ wave, to random heterogeneity in the lithosphere beneath single stations or local arrays. The first such study (Aki 1973) used Chernov (1960) theory to explain amplitude and time delay variation of 2-s period $P$ waves beneath the Montana Large-Aperture Seismic Array (LASA). In Chernov theory, heterogeneity has a Gaussian autocorrelation function characterized by a root-mean-square (RMS) velocity fluctuation $\varepsilon^2$ and correlation length-scale $a$:

$$\langle v(x)v(x + H) \rangle = \varepsilon^2 \exp \left( -\frac{|H|^2}{a^2} \right)$$

(1)

where the angle brackets imply integration over $x$. 
Figure 1  Record section of an earthquake in the Tonga-Fiji subduction complex (January 17, 1995; \(M = 6.0\); hypocentral depth of 637 km) recorded by the Hawaiian Volcano Observatory seismic network. Almost all of the energy in the slowly decaying and spatially incoherent coda following the impulsive, short-duration \(P\)-wave is generated by scattering in the volcanic edifice and lithosphere below the array.

For LASA, a layer extending to \(\sim 60\) km depth with \(\varepsilon^2 = 4\%\) and \(a = 10\) km explains the data well (Aki 1973). Other authors report longer correlation lengths for the Norwegian Seismic Array \((a = 15-60\) km\) (Berteussen et al 1975), the southern California seismic array \((a = 25\) km\) (Powell & Meltzer 1984), and the Gauribidanaur array in southern India \((a = 20\) km\) (Berteussen et al 1977), indicating variation in the fundamental scale of lithospheric structure beneath the four sites. However, the single-layer, Gaussian autocorrelation representation of random inhomogeneity in these studies is extremely limited by poor data resolution. Flatté & Wu (1988) introduced a new measure of amplitude and phase coherence, enabling them to significantly expand the representation of lithospheric heterogeneity. Working with Norwegian Seismic Array (NORSAR) data, they obtained a multilayer power law model of compressional velocity fluctuation, extending to \(\sim 250\) km depth (Flatté et al 1991). In the upper crust, above \(\sim 15\) km depth, the heterogeneity spectrum is white between 5.5 and 110 km wavelength (the spectrum is unconstrained outside
this band). Between 15 and 175 km depth, the lithosphere is enriched in long-wavelength variability while retaining greater short-wavelength variation than the subjacent asthenosphere. This is indicative of diffusive processes erasing shorter-scale-length heterogeneity in the asthenosphere. Retention of significant short-wavelength variability in the lithosphere is consistent with the work of Silver and co-workers (e.g. Silver & Chan 1991), documenting whole lithospheric strain fabrics persisting over hundreds of millions of years. Clearly, the Mohorovicic discontinuity is not the base of geology (Silver 1996).

The Russian peaceful nuclear explosion refraction profiles document a $Pn$ wave propagating within the shallow subcrustal lithosphere to distances of several thousand kilometers (Ryberg et al 1995) and followed by a high-frequency coda. Conventional layered Earth models do not predict $Pn$ at these distances, nor do they produce significant coda energy. Tittgemeyer et al (1996), however, show that multiple scattering within a highly heterogeneous mantle lid produces teleseismic $Pn$ phases and high-frequency coda. Superposition of velocity fluctuations with $\varepsilon^2 = 4\%$ and a vertical correlation length scale of 2 km onto a positive upper mantle velocity gradient reduces the effective velocity of the upper mantle $P$ wave at low frequencies and results in multiple scattering of the higher-frequency components. This highly heterogeneous waveguide must extend to at least 100 km depth. The short correlation length scale may be an artifact of modeling, which restricted velocity fluctuations to one dimension, but it is more likely a true property of cold Russian platform lithosphere reflecting differences in the correlation length scale of heterogeneity in the horizontal (long) and vertical (short) directions (e.g. Wagner 1996, 1997).

Although the long-range Russian profiles are unique, European controlled source profiles display similar levels of $Pn$ coda and there are abundant single-station observations of $Sn$ and $Pn$ at teleseismic distances (Molnar & Oliver 1969). Enderle et al (1997) take this as evidence of an extensive upper mantle $Pn$ waveguide. What is responsible for the velocity fluctuations is a question of considerable interest. Without access to lithospheric mantle it is difficult to associate scale lengths and RMS velocity fluctuations to specific geologic structures or the causative processes. In recent years, there have been several attempts to image localized scatterers in continental lithosphere by using local array recordings of regional and teleseismic waves. These studies use a hybrid parameterization of heterogeneity, specifying scattering strength (or some associated parameter) as a function of position. We refer collectively to the resulting techniques as scattered-wave imaging. The two most common approaches, semblance (Lay 1987, Lynnes & Lay 1989) and stochastic Kirchhoff migration (Revenaugh 1995b), both rely on travel time information to map arrivals in coda to specific scatterer positions but differ in how they detect these arrivals. Semblance uses a measure of waveform similarity that assumes
scattered waves radiate in all directions with equal magnitude and unchanging polarity. Stochastic Kirchhoff migration uses fluctuations of coda energy, rather than waveform similarity, eliminating the constraint on scattered wave polarity. By design, both are better able to detect isolated features with scale lengths comparable to seismic wavelength ("point" scatterers) than extended bodies or surfaces. Neither is capable of estimating the absolute strength of scattering; each provides a measure of relative variation.

Using the semblance method, Mohan & Rai (1992) image a strong scatterer in the lower crust and upper mantle west of the Gauribidanur (India) array. They associate this feature, localized between 35 and 55 km depth, with a granitic intrusion thought to be a Precambrian suture zone between the east and west Dharwar craton.

Revenaugh (1995b) uses stochastic Kirchhoff migration to image low-angle forward scatterers in the lithosphere beneath southern California. As applied to a data set of short-period recordings of teleseismic $P$ waves, the method reveals a “curtain” of high scattering strength paralleling the southern flank of the Transverse Ranges (Figure 2). This feature coincides with a slab of

![Figure 2](image)

*Figure 2* Image of scattering strength for a depth slice at 150 km beneath southern California. An east-striking sheet of strong scattering dips almost vertically to greater than 250 km depth. This marks the southern flank of a high-velocity slab that has been independently imaged by seismic tomography.
high $P$-wave velocity in the upper mantle seen in tomographic images (e.g. Humphreys & Clayton 1990). Associated with subcrustal lithospheric subduction of both the North American and Pacific plates through the Big Bend region of the San Andreas fault system, this primarily thermal anomaly should not be an efficient scatterer of short-period ($\sim 1$ s) $P$ waves. Revenaugh (1995b) reconciles this discrepancy by appeal to progressive tearing of the downgoing slab, juxtaposing cool slab interior with warmer mantle rocks. Localized highs in the scattering pattern underlie along-strike steps in the Transverse Ranges, suggesting a dynamic tie between slab tears and shallower orogenic processes. Away from this structure, scattering variability in the upper mantle is minimal. Using a similar technique applied to regional $S$-wave coda, Nishigami (1991) also finds low levels of lateral variation in upper-mantle scattering strength.

REFLECTIVITY OF CONTINENTAL CRUST

The deterministic model of Earth structure, in particular splitting of the heterogeneity spectrum into long-wavelength background variations and short-wavelength reflectors, was developed to model the subhorizontal layering and strong impedance contrasts of shallow, sedimentary sections. Seismic reflection profiling is premised on this description of the subsurface, beginning with acquisition and continuing through to processing and interpretation. Its success in exploration is a testament to the power of the deterministic description in the shallow environment. As applied to crystalline basement rocks, however, this description is a signal failure. Laterally coherent reflectors are almost entirely absent, replaced by short, anastomosing reflections and a much more diffuse “coda.” Common processing and interpretation techniques assuming single scattering and a horizontal heterogeneity spectrum depleted in high wavenumber energy produce false coherence, biasing the result toward the deterministic model (Gibson & Levander 1990, Emmerlich et al 1993). Progress in this environment requires three things: ($a$) a statistical model of the subsurface capable of explaining relevant aspects of the backscattered wave field, ($b$) sufficient theory and numerical ability to accurately model wave propagation, and ($c$) the
means to robustly extract information from data. In a series of papers, Levander, Holliger, Goff, and other workers have addressed these three fronts.

Initial development of a statistical model of crystalline crust relied on models of continuous Gaussian or self-similar media, similar to those used in studies of subcrustal lithosphere. Although it is simple to use, continuous inhomogeneity does a poor job of representing exposures of former mid- and lower-crustal rocks that tend to be structurally complex assemblages of distinct lithology, rather than continuously variable fields. This discrepancy led to the introduction of “modal” fields: two- and three-dimensional fields that assume only a small number of discrete values or lithologies (Goff et al 1994).

The spatial variation of modal fields is described by an autocorrelation function. This is typically the von Karman autocorrelation function developed for turbulent media, which describes a self-similar or self-affine fractal field with variability at all length scales and no periodicities (a self-affine field requires an anisotropic rescaling to appear self similar). The von Karman autocorrelation constrains only the amplitude spectrum of the heterogeneity, saying nothing about phase, emphasizing its inherently statistical nature. The success of this description in modeling field observations of exposed basement rocks (Holliger et al 1993, Levander et al 1994) suggests that the modal field description is a sufficient characterization of the deep seismic structure of continental crust.

Unless the RMS variation of seismic heterogeneity is small (several percent or less), multiple scattering will be important. The effects of data processing steps assuming dominant single scattering, or predictable multiple scattering as in the case of reflection “multiples,” must be understood. These effects are manifest both in deterministic imaging of the subsurface (e.g. migration) and in statistical inference. For example, one must understand the effects of tuning and detuning of seismic waves by various length scales of heterogeneity before attempting to interpret amplitudes in terms of RMS velocity variation.

Much work has gone into the analytical description of multiply scattered wave fields in inhomogeneous media. For elastic (vector) waves this is enormously difficult, but progress is being made (e.g. Li & Hudson 1995, 1996; Shapiro & Treitel 1997). At present, the most efficient means of understanding wave interaction with random inhomogeneity is through Monte Carlo modeling (e.g. Wagner & Langston 1992, Wagner 1996). A number of investigators have followed this approach, investigating the effects of standard seismic processing on synthetic data for random, two-dimensional acoustic heterogeneity. As a summary of the primary results, I note that (a) common-midpoint stacking, acting as a dip filter, artificially enhances the horizontal coherency of images of random inhomogeneity (Gibson 1991); (b) migration does not eliminate this smearing and produces erroneous horizontal coherence, even for a scattering medium with little or no lateral continuity (Emmerlich et al 1993); and (c) unmigrated
images bear little resemblance to the underlying structure and individual reflection sequences may represent interference more than structure (Holliger et al 1994). These effects worsen as the importance of multiple scattering increases. To date, most active-source studies have synthesized data in only two spatial dimensions, have coupled \(P\)-wave velocity with density, and have ignored \(S\)-wave velocity variability. Although the latter appears to be relatively unimportant for near-vertical incidence (Emmerlich 1993), it is very likely that these studies underestimate the amount of image distortion induced by multiple scattering. More powerful prestack migration and three-dimensional imaging are likely to improve the situation, but these expensive techniques find only rare application in deep crustal imaging where the deleterious effects of multiple scattering are worst.

This somewhat grim reality beckons the question: what information can we meaningfully extract from deep crustal reflection profiles? The answer, of course, depends on the setting, both geological and experimental. Deterministic structure, which satisfies the single scattering approximation, may be evaluated, although the focusing and defocusing effects of random inhomogeneities will introduce amplitude and phase distortions. Of the latter, only a statistical description is possible. The aspects of a statistical description that can be determined depend on the inhomogeneity itself, the accuracy of the assumed description, and the temporal and spatial resolution of the data. For example, Gibson (1991) relates the lateral coherence of active-source reflection data to lateral coherence of the inhomogeneity for a smooth (nonmodal) field with little multiple scattering. On the other hand, Holliger et al (1992) show that no similar estimate is possible for a model of deep crustal inhomogeneity based on the Ivrea Zone. In this case, homogenization of the wave field due to multiple scattering largely desensitizes the wave field to lateral coherence of the medium, and it is difficult to relate the remaining sensitivity to true lateral coherence of the medium (Hurich 1996).

Pullammanappallil et al (1997) discuss a Monte Carlo based optimization scheme for estimating the horizontal characteristic length scale \(a_x\) and Hurst number \(\nu\) (a parameter of the von Karman autocorrelation related to fractal dimension) from seismic exploration data. They apply this method to two deep crustal seismic profiles in the Basin and Range Province. For depths of 12 to 21 km, \(a_x\) is \(\sim 240\) m and \(\nu = 0.5\) (fractal dimension of \(\sim 2.8\)). The latter is equivalent to the value determined from field studies of the Ivrea Zone, Italy (Holliger et al 1993) and upper- and middle-crustal rocks of the Lewisian gneiss complex, Scotland (Levander et al 1994), but much larger than values for the shallow upper crust where small cracks dominate the scattering inhomogeneity (Holliger 1996, 1997; Leary & Abercrombie 1994). The characteristic length scale is roughly a quarter of the first Fresnel zone at depth and much shorter
than the length scales estimated for the subcrustal lithosphere. Along these
lines, there is a clear break in reflectivity structure (e.g. Mooney & Meissner
1992, Enderle et al 1997) and a change in scattering character (Ritter et al 1997)
associated with the Mohorovicic discontinuity, suggesting different length scales
of heterogeneity above and below. As estimates of this type become more com-
mon, it should be possible to infer dominant processes involved in the production
of inhomogeneity.

SCATTERING AND SEISMOGENESIS

There is a growing body of literature suggesting that structural features within
the crust strongly influence the seismogenic process. Some of the best indica-
tions draw from travel time tomography, showing clear correlations of velocity
anomalies with centers of main shock slip and aftershocks (e.g. Michael &
not seem surprising—certainly, crustal properties modulate slip, and some, if
not much, of the richness of the distribution of seismicity must mirror com-
plexities of the crust. However, numerical simulations of rupture on interacting
faults show that it is quite possible to produce complex spatial and temporal
patterns in a uniform crust (e.g. Rundle 1988). To what extent and in what
ways is the seismogenic process influenced by properties of the surrounding
crust? Along with comprehensive seismic and geodetic monitoring, an answer
to this question requires that we thoroughly characterize the crust around active
faults. This entails many geological and geophysical probes. Seismic scattering
figures heavily.

An early observation in the study of regional S-wave coda is the pronounced
variation in coda $Q^{-1}$ between tectonically active and stable areas. Coda $Q^{-1}$,
a measure of the attenuation rate of coda-wave amplitude with time, is highest
(attenuation greatest) in active regions (e.g. Singh & Hermann 1983). Coda
$Q^{-1}$ reflects a complex mixture of intrinsic and scattering attenuation, the latter
measuring the rate at which energy is scattered out of a wave front. Geographic
variability in coda $Q^{-1}$ is due largely to variation in intrinsic attenuation, which
is greater in tectonically active (warm) regions. However, scattering attenuation
also varies and is greatest in extensively faulted regions (e.g. Nishigami 1991,
Jin et al 1994). This implies that the crust, and perhaps the uppermost mantle,
surrounding active faults is highly “opaque” to seismic waves, that is, that it
scatters seismic waves strongly.

Coda attenuation measurements sample a roughly ellipsoidal volume of crust
and upper mantle containing the source and receiver. To minimize the impact of
nonisotropic scattering and nonuniformly distributed scatterers, measurements
are made at long lapse times and thus average over a large area. This averaging
clouds the connection between scattering and faults, in particular, the relative roles of the fault zone versus distributed cracks.

Several recent studies attempt to constrain more tightly the geographic distribution of scattering strength proximal to faults. Nishigami (1991, 1997) uses the single scattering model of Sato (1977) as the basis of an inversion method for imaging variations of volumetric scattering strength in three dimensions. For regional $S$-wave coda data from the Hokuriku district, central Japan, scattering strength in the upper crust near mapped faults is 20% greater than regional background levels. Using power scaling rules (Aki and Richards 1980), Nishigami concludes that velocity fluctuations near the faults are 10% greater than elsewhere. The scattering highs do not clearly define the fault traces, however, and are spread over 10–30 km, suggesting that extensive cracking outside the fault zone proper may contribute significantly to scattering.

Matsumoto et al (1998) apply a migration-like operator to active-source reflection data sampling the source region of the 1995 Kobe, Japan ($M = 7.2$) earthquake. Their images of subsurface $P$-wave scattering strength reveal two patches of strong scattering. One, spread out between 10 and 25 km depth, lies southwest of the hypocenter in a complexly faulted region. The second lies just below the hypocenter, suggesting that rupture initiated in or near a highly inhomogeneous region characterized by structural scale lengths of order 500 m. The tomographic study of Zhao et al (1996) identifies the hypocenter as a low $P$-wave velocity, high Poisson’s ratio region, which they attribute to a local concentration of fluid-filled cracks, a natural candidate for strong scattering.

Revenaugh (1995c) modified the stochastic Kirchhoff migration method of Revenaugh (1995b) to image scattering variations in the crust. This approach has the considerable advantage of not relying on local seismicity, allowing imaging of quiescent fault segments. As implemented, it estimates geographic variation of the statistical significance of scattering (or scattering “potential”) in the upper crust (<12–15 km depth) rather than the strength of scattering, a much less stable quantity. Scattering potential is a measure of the likelihood that scattering strength locally exceeds the regional mean; scattering potentials near unity imply locally strong scattering, whereas potentials near zero mark the weakest local scattering strengths. In other words, scattering potential is a relative index of scattering strength.

Color Plate 1 shows scattering potential for the region of southern California surrounding the June 28, 1992, Landers earthquake ($M_W = 7.3$). Scattering potential varies considerably over a variety of length scales. The highest scattering potentials occur close to the rupture area, but they do not delineate the fault, and there are high scattering potential patches far from mapped fault segments. There is, however, a noticeable tendency for aftershocks to cluster in regions of high scattering potential. This observation is clearer in Figure 3,
which compares along-fault variation in scattering potential and aftershock density. There are a number of implications of this clustering. Foremost, it requires that aftershock occurrence be tied, at least in part, to properties of the crust near the fault. This raises a question: Does repeated slip on the fault and/or near-fault stress variation induce scattering by increasing fracture density, for example, or does crustal structure predating fault development influence present-day seismogenic processes? The answer may lie anywhere in between these two extremes, but the correlation of a measurable property of the crust with aftershock density and coseismic slip (Revenaugh 1995c) implies some predictability of the earthquake process, at least in terms of along-fault slip variability.

To further elucidate the relation of crustal scattering and seismogenesis, Revenaugh & Reasoner (1997) image the San Andreas fault system through central and southern California by using stochastic Kirchhoff migration. The paucity of seismicity along many stretches of the San Andreas precludes meaningful correlation of scattering potential and aftershock seismicity. At present, the instrumental record of seismicity is too short to allow such a comparison. Revenaugh (1995c) notes that transitions from high to low scattering correlate geographically with fault jumps and bends along the Landers rupture zone, prompting Revenaugh & Reasoner (1997) to look for a similar correlation between segment bounds and scattering along the San Andreas.

Segment bounds delimit fault stretches that rupture singly or multiply in large earthquakes. They are determined from historic and paleoseismicity and geologic constraints, making them more representative of long-term fault behavior than the short instrumental record alone. Revenaugh & Reasoner (1997)
Figure 4  Along-fault gradient in scattering potential and segment bounds (vertical bars) for a ∼500 km stretch of the San Andreas Fault in central California. Six of nine segment bounds (Working Group on California Earthquake Probabilities 1988) lie in regions of high scattering gradient (gradients outside the one standard deviation level), an outcome with a <5% chance of random occurrence.

find that segment bounds coincide with high along-fault scattering gradients for the San Andreas fault in central California (Figure 4). The correlation of segment bounds and scattering gradients suggests either that segment bounds delimit distinct crustal blocks (with distinct scattering signatures) or that segment bounds mark transitions between distinct stress states that dynamically influence scattering, the latter potentially manifested as changes in crack density or crack orientation. Distinguishing these two hypotheses is the same question of precedence raised by Revenaugh (1995c) for scattering near the Landers earthquake. Revenaugh & Reasoner (1997) consider it in unique fashion, by searching for evidence of offset in the pattern of scattering potential along the San Andreas fault. They obtain optimal cross-fault correlation for 315 km of right-lateral offset (Figure 5), a value in excellent agreement with geologic estimates of cumulative offset. The fact that scattering offsets (that is, that some portion of the scattering field advects with the crust), suggests that preexisting structure imparts a strong influence on fault structure.

Temporal Change
The static offset of scattering potential does not exclude a dynamic component. Indeed, there is mounting evidence of a component of crustal and shallow
mantle scattering that responds dynamically to changes in the regional stress field. This comes from temporal variation in coda $Q^{-1}$ of regional S waves.

Intrinsic attenuation is an essentially static property of the shallow lithosphere, such that any temporal variation in coda $Q^{-1}$ must reflect changes in scattering attenuation. Significant temporal variation in coda $Q^{-1}$ has been observed in several tectonic settings and over time scales ranging from weeks to decades (Chouet 1979, Jin & Aki 1989, Su & Aki 1990). More important is the apparent connection with seismicity: large temporal gradients in coda $Q^{-1}$ often coincide with changes in seismic $b$ value, a measure of the production rate of small earthquakes relative to large (e.g. Jin & Aki 1989). However, the sense of correlation is not consistent—an increase in $b$ value may accompany an increase or a decrease in coda $Q^{-1}$—and there are coda $Q^{-1}$ gradients with no obvious association to seismicity. Jin & Aki (1989) and Aki (1992) suggest that this behavior indicates changes in crack density in the semiductile lithosphere. Scattering strength and attenuation are highly sensitive to crack density, causing coda $Q^{-1}$ to vary. The sense of variation, increase or decrease, depends on the scaling of dominant seismic wavelength to fracture size. Likewise, small earthquakes on these fractures may raise or lower seismic $b$ value depending on the local stress state and the method of computing $b$ value. The result is a coincidence in timing, but no consistency in sign. The spatial and temporal coarseness of coda $Q^{-1}$ and seismic $b$-value measurements make it difficult to envision an effective monitoring program based on these observations, but the correlation is intriguing and may offer insight into the boundary conditions of the seismogenic process.

Figure 5  Cross-fault comparison of scattering potential along the San Andreas Fault in central California after removal of 315 km of right-lateral offset. Profiles have been filtered to remove long-wavelength variations. The excellent correlation implies scattering advects with the crust. Its relationship to aftershock seismicity (Figure 4) and segment bounds (Figure 5) implies seismogenesis is strongly influenced by heterogeneity predating the fault.
Other Scattered-Wave Imaging Studies
Chávez-Pérez & Louie (1998) use regional P-wave coda to investigate crustal structure near the 1994 Northridge earthquake. Combining aspects of controlled-source profiling and scattered-wave imaging, they map a north-dipping reflector, possibly corresponding to the Elysian Park Thrust (Davis & Namson 1994), and an offset subhorizontal reflector that they interpret as a midcrustal detachment. Innovative studies of this sort have the potential to greatly advance our knowledge of the deep configuration of faults mapped in the near surface.

Scattered waves are an important component of earthquake strong motion. Spudich & Iida (1993) use a source array to image scatterers contributing to the early S-wave coda. They find that laterally propagating waves scattered from a basin edge were prominent in the early coda. These are a mixture of S and surface waves \( P_g \). A number of other studies note basin-edge surface waves (e.g. Frankel & Vidale 1992), which appear to be a dominant component of early coda. Revenaugh (1995a) and Revenaugh & Mendoza (1996) demonstrate a regional correlation of surface topographic roughness and the efficiency of \( P \) to \( R_g \) scattering, which they use to formulate a parametric model of high-frequency \( R_g \) velocity. Knowledge of \( R_g \) producers (scatterers) and propagation velocity will vastly improve seismologists’ ability to predict ground motion.

SCATTERING AND VOLCANOES
Seismic scattering is finding increasing application as a means of quantifying the short scale-length heterogeneity and magma plumbing of active volcanoes. Scattering is well suited to this task. First arrivals avoid low-velocity magma bodies, which are difficult to image tomographically unless they are large, but can be very strong scatterers. Likewise, faults and thin layers have little effect on travel time, but they do produce coda arrivals.

Nishigami (1997) applies his coda-envelope inversion (Nishigami 1991) to two active volcanoes in central Japan (Mount Ontake and Mount Nikko-Shirane). Strong scattering beneath Mount Ontake extends to a depth of \( \sim 7 \) km, close to the maximum depth of microseismic activity. Strong scattering is confined to a narrow linear zone to 7 km depth, where it spreads out laterally. Interestingly, the zone of high scattering is nearly void of microseisms. A similar line of high scattering strength extends beneath Mount Nikko-Shirane to 20 km depth. In both cases the coincidence of the volcano and the vertical zone of high scattering suggests that a magma conduit is responsible.

Mikada et al (1997) adapt a diffraction tomography approach to image scattering beneath Izu-Oshima volcano, an island about 100 km south of Tokyo, Japan. Their detailed image shows several patches of strong scattering. Most conspicuous is a cloud of scatterers centered at \( \sim 9 \) km depth directly beneath
the volcano crater. Mikada et al associate this feature with the primary magma reservoir. They further associate smaller and shallower patches of high scattering strength with subchambers tapped during eruptive activity in 1986.

Nishimura et al (1997) apply an envelope inversion method to active-source recordings from the Jemez volcanic field, New Mexico. Their profiles of the depth dependence of scattering peak in the upper 5–7 km beneath the Valles Caldera at a level more than a factor of five greater than similar profiles for the Rio Grande Rift and Colorado Plateau. They conclude that two episodes of eruption and collapse of the caldera, the presence of rhyolite domes, pyroclastic material, and intrusions into caldera fill are responsible. Scattering beneath the Rio Grande Rift is stronger than that beneath the Colorado Plateau, again indicating a significant role of intrusion and volcanism in the production of scattering heterogeneity.

CONCLUSIONS

For many years, seismology had one way of parameterizing the Earth: as a layered structure colored by smooth velocity perturbations, much like a Mark Rothko painting except that the edges are sharp (reflectors). Capturing much of Earth’s inhomogeneity, this representation has been enormously successful, but seismology, in particular, seismic scattering, has much more to tell us. Scattered waves feel the “fabric” of crust and mantle rock and carry that information to the surface. The pictures they paint vary with the style of representation. The purely statistical representation is like a Jackson Pollock drip painting: what matters is the fabric, not the individual arcs of color. To realize its full potential, the sophistication of representation and modeling must rise, giving full consideration to vector waves, realistic background structures, and three-dimensional wave propagation, as well as the connection between scattering heterogeneity and geology. The work of Levander, Holliger, and others demonstrates the tremendous promise of using geology to design statistical representations of seismic inhomogeneity, but there is much work left to do. The same is true of scattered-wave imaging, which sees the world through the eyes of a pointillist, as a distribution of individual scatterers. To continue making progress, the relation of scatterers to geology and other geophysical images of the subsurface must be better understood, and seismology must strive to make the scatterers more versatile, better able to capture the essence of complex larger structure. In these regards, scattered-wave seismology is still on the steep limb of the learning curve.

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Color Plate 1  Scattering potential of the upper crust surrounding the 1992 Landers earthquake (epicenter marked by bull’s-eye). ML ≥ 2.0 aftershock seismicity (black dots) clusters in areas of highest scattering potential. Green and black line marks fault line used in Figure 4.