Properties of seismic fault zone waves and their utility for imaging low-velocity structures

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Abstract. A two-dimensional solution for the scalar wave equation in a model of two vertical layers between two quarter spaces is used to study properties of seismic waves in a laterally heterogeneous low-velocity structure. The waves, referred to as seismic fault zone waves, include head waves, internal fault zone reflections, and trapped waves. The analysis aims to clarify the dependency of the waves on media velocities, media attenuation coefficients, layer widths, and source-receiver geometry. Additional calculations with extreme low-velocity layers provide examples that may be relevant for volcanic and geothermal domains. The interference patterns controlling seismic fault zone waves change with the number of internal reflections in the low-velocity structure. This number increases with propagation distance along the structure, decreases with fault zone width, and increases (for given length scales) with the velocity contrast. The relative lateral position of the source within the low-velocity layer modifies the length scales associated with internal reflections and influences the resulting interference pattern. Low values of Q affect considerably the dominant period and overall duration of the waves. Thus there are significant tradeoffs between propagation distance along the structure, fault zone width, velocity contrast, source location within the fault zone, and Q. The lateral and depth receiver coordinates determine the particular point where the interference pattern is sampled and observed motion is a strong function of both coordinates. The zone connecting sources generating fault zone waves and observation points with appreciable wave amplitude can be over an order of magnitude larger than the fault zone width. Calculations for cases with layer P wave velocity of ~200 m s⁻¹, modeling a vertical dike or crack with fluid and gas, show conspicuous persisting oscillations. The results resemble aspects of seismic data in volcanic domains. If these waves exist in observed records, their explicit recognition and modeling will help to separate source and structural effects and aid in the interpretation of volcano-seismology signals. Although the tradeoffs in parameters governing seismic fault zone waves are significant, each variable has its own signature, and the parameters may be constrained by additional geophysical data. Simultaneous modeling of many waveforms with an appropriate solution can resolve the various parameters and provide a high-resolution structural image.

1. Introduction

The definition and description of a fault zone (FZ) depends on the application at hand, imaging method, and location. In some places, detailed mapping of surface features [Johnson et al., 1997] documents earthquake rupture zones that consist of shear bands 50-500 m wide. At the other extreme, field and laboratory data indicate that the core of the San Gabriel FZ is only a few centimeters wide [Evans and Chester, 1995]. These geological studies provide direct evidence on FZ properties, but they are limited to shallow depth. The imaging of deep crustal structures requires inversion of seismological and other indirect data. An accurate determination of FZ properties at depth provides a preliminary basis for a variety of studies ranging from routine derivation of earthquake parameters (e.g., location and focal mechanism) to mechanical simulations of crustal responses (e.g., dynamic rupture histories, pre-earthquake and post-earthquake deformational fields).

At the Parkfield section of the San Andreas fault (SAF) a borehole seismic network enables tomographic inversions to constrain a fairly complex local crustal velocity structure [Lees and Malin, 1990; Michelini and McEvilly, 1991; Michael and Eberhart-Phillips, 1991; Ben-Zion et al., 1992]. The inverted velocity structure varies transversely to and along the strike of the SAF. The FZ structure, roughly defined by a 2-3 km wide band of seismicity along the strike of the fault, corresponds to a broad transition zone in P and S wave velocities [e.g., Michelini and McEvilly, 1991]. Such broad FZ structure is seen elsewhere in central California [e.g., Feng and McEvilly 1983; Steierman, 1984; Mooney and Ginzburg 1986; Eberhart-Phillips et al., 1995]. In each of these studies, however, the best available seismic velocity model is incapable of resolving the actual surface or volume sustaining the discontinuous motion of faulting. As mentioned above, where surface fault breaks are observed, earthquake deformation is found to be concentrated in strips that are a few hundred meters wide or less. Outside these narrow slip zones there may exist extensive widths of previously damaged low-velocity rock, and such broad deformation zones are probably the objects imaged by the usual seismic, electrical, and gravity surveys. However, studies of the 2-3 km cumulative damage

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zones do not resolve details associated with the preparation and rupture processes of individual seismic failure episodes.

Highly damaged (gouge) FZ layers and other low-velocity structures with systematic material interfaces, such as molten dikes, fluid/vapor-filled cracks in geothermal areas, coal seams, etc., can generate specific seismic phases indicative of their velocity structure. For brevity we use below the term "fault zone" to denote the class of such layered low-velocity structures. Seismic waves generated by a FZ structure can be used to image properties of the low-velocity structure at depth. Understanding and using such data require detailed modeling computations. Cormier and Spudich [1984], Hori et al. [1985], and Cormier and Beroza [1987] performed ray-tracing calculations in models incorporating low-velocity FZ layers. Ben-Zion [1989, 1990] and Ben-Zion and Aki [1990] derived analytical solutions for waves radiated by sources in laterally heterogeneous structures containing vertical plane-parallel interfaces. The solutions of Ben-Zion [1989] and Ben-Zion [1990] involved two-dimensional (2-D) and three-dimensional (3-D) problems with a single interface, respectively. Ben-Zion and Aki [1990] obtained a 2-D solution for scalar (i.e., acoustic P or antipole S) elastodynamic waves in a half space with an arbitrary number of vertical layers.

When used in conjunction with P wave velocities, the solution of Ben-Zion and Aki [1990] models near-structure P waveforms with P-to-P fault zone head waves (FZHWs), direct P waves, and multiply reflected P waves propagating along the low-velocity structure. When the solution is assigned typical S wave velocities, it models near-structure shear waveforms with S-to-S FZHWs, direct S body waves, and shear waves that are internally reflected in the FZ. For certain velocity structures and source-receiver configurations the internally reflected (P or S) waves produce prominent interference patterns called fault zone trapped waves (FZTWs).

FZHWs are seismic disturbances that propagate along material interfaces in the FZ with the velocity and motion polarity of the body waves on the faster side of the interface. These phases are analogous to head waves in horizontally layered media. FZHWs arrive at near-structure stations in the slower velocity medium before the direct body waves, and they are characterized by an emergent waveform having reversed first-motion polarity from that of the direct arrival. Misidentification of FZHWs as direct arrivals can introduce first-motion polarity from that of the direct arrival. This and other differences may be reduced by using more complex source functions or a superposition of sources in different locations. The latter is illustrated with example calculations, although we emphasize that our goal is not to produce realistic volcanic seismograms but rather to point out the possible existence of FZ waves in volcanic seismic data.

Another goal of the paper is to discuss a possible application of seismic FZ waves in the context of volcano and geothermal seismology. Using the solution of Ben-Zion and Aki [1990] with model properties corresponding to an extreme low-velocity layer (and a simple dislocation source), we generate seismograms for an idealized structure of a vertical dike or crack with fluid and gas in a half space. The synthetic seismograms for such cases show long-duration wave trains reminiscent of volcanic seismic records. The results are, however, different in several important aspects from typical volcanic seismograms. In particular, the calculated time histories begin with a large-amplitude signal, while the amplitude of observed volcanic seismograms usually builds up gradually. This and other differences may be reduced by using more complex source functions or a superposition of sources in different locations. The latter is illustrated with example calculations, although we emphasize that our goal is not to produce realistic volcanic seismograms but rather to point out the possible existence of FZ waves in volcanic seismic data.
The calculations of this work show that seismic FZ waves are sensitive to a large set of parameters. The high sensitivity of the waveforms allows small details of structure and source-receiver geometry to be resolved. However, the strong coupled dependency of the waveforms on a set of parameters requires careful observational and theoretical analyses. The results indicate that observational work should consider volumes in a parameter space including propagation distance along the structure, FZ width, velocity contrast, source location within the FZ, and \( Q \). Simultaneous modeling of a large data set containing seismic FZ waves with an appropriate solution can provide high resolution images of coherent low-velocity layers in important geological environments.

2. Model

The model forming the basis for the calculations done in this paper consists of two quarter spaces separated by two vertical FZ layers (Figure 1). All variables are independent of the \( y \) direction along strike. The media are numbered sequentially from left to right. An \( SH \) line dislocation is located at a general \((x_s, z_s)\) position in medium 2. The source can be placed inside or outside the FZ by setting the parameters of medium 2 different from or equal to those of medium 1. The receiver is located at a general \((x, z)\) position. Thus both source and receiver can be moved up and down and in and out of the FZ.

From Ben-Zion and Aki [1990] the displacement field in the four media model of Figure 1, satisfying the equation of motion and stress-displacement boundary conditions, can be written as

\[
 V_1(x,z,t) = \frac{1}{2\pi} \int_{-\infty}^{\infty} \int_{-\infty}^{\infty} B_1'((k,\omega)e^{+y_s x} \cos(\omega z)e^{+i\omega t} dk d\omega
\]

\[
 V_2(x,z,t) = \frac{1}{2\pi} \int_{-\infty}^{\infty} \int_{-\infty}^{\infty} [B_2'(k,\omega) + h(x, -x) B_1'(k,\omega)]e^{+y_s x} e^{+i\omega t} dk d\omega
\]

\[
 V_3(x,z,t) = \frac{1}{2\pi} \int_{-\infty}^{\infty} \int_{-\infty}^{\infty} B_3'(k,\omega)e^{+y_s x} \cos(\omega z)e^{-i\omega t} dk d\omega
\]

\[
 V_4(x,z,t) = \frac{1}{2\pi} \int_{-\infty}^{\infty} \int_{-\infty}^{\infty} B_4'(k,\omega)e^{-y_s x} \cos(\omega z)e^{-i\omega t} dk d\omega
\]

where \( k \) is vertical wavenumber, \( \omega \) is angular frequency, \( c_j \) are wave velocities (\( P \) or \( S \) depending on the application) in medium \( j \), \( \gamma_j \) is the impedance of medium \( j \), and \( D \) is the magnitude of the dislocation (a constant 1 cm in this work). The other reflection/transmission coefficients are given by

\[
 B_1' = 1/(2D) \{ [F_1' + F_2'G_1' - F_1'G_1'] B_1' + G_1'F_1' \}
\]

\[
 B_2' = 1/(2D) \{ -F_1'G_1' + F_1'G_1' \}
\]

\[
 B_3' = 1/(2D) \{ -F_2'G_1' + F_2'G_1' \}
\]

\[
 B_4' = 1/(2D) \{ F_2'G_1' - F_2'G_1' \}
\]

and \( W_2 \) and \( W_3 \) are the widths of media 2 and 3, respectively.

Using discrete wavenumber summation for the \( k \) integrals and fast Fourier transform for the \( \omega \) integrals, the set (1) can be evaluated for various model parameters. Attenuation is incorporated into the solution by multiplying the wave velocities \( c_j \) in (1) with the function \([1 + \ln(\omega/2\pi)/(\pi Q_j)] - i(2\pi j)\), where \( Q_j \) are quality factors for medium \( j \). Instrument response and various source functions can be implemented by appropriate multiplications in the spectral domain. For computer programming it is useful to group together various exponential functions in (1) so that the arguments of the exponents are small during calculations. In the applications below, the source is characterized by a unit step function \( h(t) \) in time and 2-D delta function \( \delta(x-x_0)\delta(z-z_0) \) in space. (Figure 15 is generated for a model with spatial periodic repetition of structure and source.) The instrument response for all calculations is flat up to a frequency of 100 Hz, and it then decreases to zero with a cosine taper that is 40 Hz wide.

3. Calculations for a Strike-Slip Fault Structure

Figure 2 shows simple reference results giving shear displacement and velocity waveforms in two degenerated versions of the model shown in Figure 1. These are (left) a homogeneous half space with shear wave velocity and quality factor \( \beta = 3.2 \text{ km s}^{-1} \) and \( Q = 1000 \), and (right) two joined quarter spaces with \( \beta = 3.2 \text{ km s}^{-1} \), \( \beta_2 = 3.0 \text{ km s}^{-1} \), and \( Q_1 = Q_2 = 1000 \). The source position is \((x_s = 0, z_s = 5.0 \text{ km})\) and receiver locations are \((x = \pm 200 \text{ m}, z = 0)\). In both cases and
Figure 1. A four-media model for a laterally heterogeneous low-velocity structure in a half space. Medium 2 contains an SH line dislocation source with coordinates \((x_s, z_s)\). Receivers marked by triangles are at general \((x, z)\) positions. Widths of media 2 and 3 are marked by \(W_2\) and \(W_3\). Quality factor, wave velocity, rigidity, and density of media \(j = 1, 4\) are denoted by \(Q_j, c_j, \mu_j, \text{and } p_j\).

In all other simulations of this paper the density of all media is \(\rho = 2.5 \text{ gr cm}^{-3}\), and the rigidity of each medium is given by \(\mu_j = \rho j^2\).

In a homogeneous half space (Figure 2 (left)) the waveforms to the left and right of the fault are the reversed images of each other, consisting of a sharp direct arrival followed by a \((t^2 - t_{\text{arr}}^2)^{-1/2}\) decay with \(t_{\text{arr}}\) denoting the arrival time of the direct wave. In the two quarter spaces model (Figure 2 (right)) the seismogram in the quarter space with the faster seismic velocity is similar to that in a half space. In the slower quarter space, however, the first arrival (for the parameters used) is an emergent head wave that propagates along the interface \(x = 0\) with the velocity and motion polarity of the faster quarter space before being radiated into the slower velocity medium [Ben-Zion, 1989]. The head wave is followed by an opposite polarity sharp direct wave traveling exclusively in the slower quarter space. The displacement fields in Figure 2 (and all figures below) approach at large time the final static deformation associated with the assumed dislocation source. Calculations with subsequent model

Figure 2. (left) Displacement and velocity seismograms in a homogeneous half space with shear wave velocity and quality factor of 3.2 km s\(^{-1}\) and 1000. In the top two seismograms the receiver position is \((x = -200 \text{ m}, z = 0)\), while in the bottom two it is \((x = 200 \text{ m}, z = 0)\). Propagation length is 5 km. (right) Corresponding results in two quarter spaces with velocities 3.2 km s\(^{-1}\) in the region \(x < 0\) and 3.0 km s\(^{-1}\) in \(x > 0\). D and H on traces denote direct and head waves.
realizations incorporating low-velocity FZ layers show much richer results than those of Figure 2.

Most of the following results in this and the next section are generated for a model with a single FZ layer. For applications of the set (1) to a model with a single low-velocity layer and an internal FZ source, media 2 and 3 are merged into one FZ layer by assigning them the same (low velocity, low rigidity, and low $Q$) parameters. For applications with a single low-velocity layer and a source external to the FZ, medium 2 is assigned the same parameters as those of medium 1, while medium 3 serves as a FZ layer.

In general, the interference patterns controlling the waveform character of the various seismic FZ waves change with the number of times the waves are reflected internally in the structure. This number increases with propagation distance along the structure, and it decreases with FZ width. Thus the primary length scale governing seismic FZ waves is the ratio of the distance the waves propagate along the structure divided by the FZ width [Ben-Zion and Aki, 1990; Igel et al., 1997]. In the present 2-D model the propagation distance along the structure is determined by the depths of the source and receiver. Thus, in applications with our 2-D model the vertical separation between source and receiver is used as a proxy for the overall propagation distance along the fault.

The critical angle of reflection within a low-velocity FZ layer, measured from the normal to the interface, decreases with the impedance contrast between the FZ and the bounding media. Thus the number of internal FZ reflections increases (for given length scales) with the velocity contrast. The relative lateral position of the source within the FZ layer modifies the length scales associated with internal FZ reflections and influences strongly the resulting overall interference pattern. Low values of $Q$ modify considerably the dominant period and overall duration of seismic FZ waves. The above theoretical considerations suggest that there are significant tradeoffs between propagation distance along the structure, FZ width, velocity contrast, source location within the FZ, and $Q$. The lateral and depth receiver coordinates determine the particular point where the interference pattern is sampled and observed motion is a strong function of both coordinates. The tradeoffs discussed above and sensitivity of the wavefield to horizontal and depth receiver coordinates are illustrated below with a set of basic simulations.

Figure 3 shows displacement and velocity seismograms in a three-media model consisting of two quarter spaces with a FZ layer in between. The media parameters are $\beta_1 = 3.2$ km s$^{-1}$, $Q_1 = 1000$, $\beta_2 = 2.0$ km s$^{-1}$, $Q_2 = 30$, $W_2 = 50$ m, $\beta_3 = \beta_4 = 3.0$ km s$^{-1}$, and $Q_3 = Q_4 = 1000$. The source operates at $(x_s = 0, z_s = 10$ km), and the receivers are at various positions along the free surface. The seismograms include FZHWs on the slower side of the fault, internally reflected phases and FZTWs in and around the FZ, and direct waves. The results demonstrate that details of seismic FZ waveforms depend strongly on the lateral receiver position. The calculations show, however, that significant amplitudes of FZ waves can exist outside the FZ proper. These effects were illustrated earlier by Ben-Zion and Aki [1990, Figures 9 and 10].

Figure 4a gives time histories and amplitude spectra of velocity seismograms for various ratios of $r/W_2$, where $r = z_s - z$ is the propagation distance along the structure. The receiver is at the free surface ($z = 0$) in the center of the FZ ($x = 0.5W_2$). The quality factor of the FZ layer is $Q_2 = 100$, and the other media parameters are the same as in Figure 3. In Figure 4a (left), $r$ is kept constant at 5 km and $W_2$ varies, while in Figure 4a (right), $W_2 = 50$ m and $r$ changes. The results demonstrate the importance of accounting properly for the
evolution of seismic FZ waves with propagation distance along the fault. As discussed above and by Igel et al. [1997], there is a strong tradeoff between \( r \) and \( W_2 \). Waveforms in a structure with a given width associated with different source-receiver configurations, and hence different values of \( r \) can exhibit changes similar to those associated with fixed \( r \) and corresponding variations of FZ widths.

An evidence of the complex interaction between the set of parameters governing seismic FZ waves is given in Figure 4b, where we show results similar to those in Figure 4a but for a more attenuative FZ having \( Q_2 = 30 \). From Figure 4b and similar calculations (see Figure 8) we find that values of \( Q \) smaller than \( \approx 50 \) modify the waveforms considerably via attenuation. Since the overall attenuation depends strongly on the propagation length, the low \( Q_2 \) value in the calculations leading to Figure 4b reverses the trend of waveform evolution with \( r/W_2 \) seen in Figure 4a.

Figure 5 compares velocity seismograms for different lateral positions \( (x_s) \) of the source within the FZ with seismograms generated by models with different FZ width. The FZ quality factor is \( Q_2 = 30 \), the source depth is \( z_s = 5 \) km, and the receiver is at \((x = 50 \text{ m}, z = 0)\). The shear wave quality factor in the FZ is \( Q_2 = 100 \).

Figure 6 (left) presents results for various receiver depths below the free surface. The media parameters are as in Figure 3, the source is at \((x_s = 0, z_s = 5 \) km\), and the lateral receiver position is \( x = 0.5W_2 = 25 \) m. Figure 6 (right) gives similar results for \((x_s = 0, z_s = 7.5 \) km\), \( W_2 = 150 \) m, and \( x = 1.5W_2 = 225 \) m. The top seismograms correspond to receiver depths of...
up to a few hundreds of meters, typical to shallow geophysical boreholes. In these cases reflections from the free surface are mixed with the other arrivals. This increases both the length and complexity of the shallow borehole waveforms, as compared to corresponding free surface seismograms. The bottom trace gives results for a receiver depth of 1.5 km. At that depth there is a complete separation between the upgoing and downgoing waveform groups for a relatively narrow FZ (Figure 6 (left)) but for a relatively wide FZ (Figure 6 (right)) those groups are still mixed. The calculations show that seismic FZ waves are very sensitive to the receiver depth. We note that the interference patterns of the waves create at given shallow depths (e.g., 175 and 250 m in Figures 6 (left) and (right)) apparent separate groups of trapped waves. Li and Leary [1990] and Li et al. [1990] interpreted two apparent groups of FZTWs in observed Parkfield borehole seismograms (station Middle Mountain; instrument depth ~250 m), using normal mode calculations for two FZ layers in a full space, as resulting from separation of fundamental and first higher modes. The results of Figure 6 indicate that such a waveform appearance can be a simple consequence of downward reflections from the free surface.

Figure 7 gives results for different velocity contrasts between the FZ and the surrounding material. Here the shear wave velocities and quality factors of media 1, 3, and 4 are set equal to 3.0 km s^{-1} and 1000, respectively. The FZ width is \( W_2 = 50 \) m, and the source and receiver coordinates are \( (x_s = 0, z_s = 5.0 \) km) and \( (x = 0.5W_2, z = 0). \) A plane \( SH \) wave with given frequency and wavenumber having a critical angle of incidence \( \theta_c = \sin^{-1}(\beta_2/\beta_1) \) will be trapped within the low-velocity FZ. The propagation distance along the low-velocity layer during one critical reflection is \( L = W_2 \tan(\theta_c) \), and the number of internal FZ reflections for given source-receiver separation along the structure is \( N = r/L \). In our case, with a line dislocation source having broad ranges of frequencies and wavenumbers, trapped waves are generated by constructive interference of internal FZ reflections having, for each frequency, values of vertical wavenumber \( k \) (common, by
Snell's law, among corresponding waves in all media) so that the horizontal wavenumber in the FZ layer ($\gamma_2$) is pure imaginary, while the horizontal wavenumbers in the quarter spaces ($\gamma_1 = \gamma_3 = \gamma_4$) are real. This situation gives an oscillatory motion in the FZ and exponentially decaying waves in the bounding quarter spaces (for more details see equations (2.6)-(2.9) and related discussion of Ben-Zion and Aki [1990]).

The number of internal FZ reflections in Figure 4a, where the velocity contrast is fixed but either the propagation path or FZ width vary, are (from top to bottom) $N = 148, 118, 99$, and $85$. Using the simple plane wave analysis discussed above as a reference guideline with $\beta = 5$ km and $W = 50$ m, we get similar values of $N$ for $\beta_2/\beta_1 = 0.56, 0.65, 0.71,$ and $0.76$. These velocity contrasts (i.e., $\beta_1 = \beta_3 = \beta_4 = 3.0$ km s$^{-1}$; $\beta_2 = 1.7, 1.9, 2.1,$ and $2.3$ km s$^{-1}$) are used in Figure 7, and comparison of the results with Figure 4a illustrates the tradeoffs between the FZ velocity contrast, FZ width, and propagation length along the fault. Although the tradeoffs are strong, they are not complete, and each parameter has its own signature. For example, the dominant period increases with increasing velocity contrast (Figure 7), increasing FZ width (Figure 4a (left)), and increasing propagation length (Figure 4a (right)). However, the overall duration of the response increases considerably with increasing velocity contrast, changes little with FZ width, and increases mildly with propagation length. Nevertheless, the results demonstrate the importance of examining jointly volumes in parameter space associated with sets of parameters before reliable conclusions on FZ properties can be made. It is also clear from the discussion above that the ratio of the velocities inside and outside the FZ, rather than individual values, is the relevant velocity parameter governing the waveforms of seismic FZ waves. Thus uncertainty in velocity of the country rock (related to typical errors of locations and origin times of earthquakes) can translate into uncertainty in the velocity of the FZ layer. Individual velocity values affect the arrival time of the phases, and these can provide important constraints for waveform inversions [Michael and Ben-Zion, 1997].

Figure 8a presents velocity and displacement seismograms for different values of $Q$ within the FZ layer. The source and receiver coordinates are $(x_s = 0, z_s = 5.0$ km) and $(x = 0.5W_2, z = 0)$. The receiver is at $(x = 100$ m, $z = 0)$ and $Q_2 = 30$. The number of internal FZ reflections in Figure 4a, where the velocity contrast is fixed but either the propagation path or FZ width vary, are (from top to bottom) $N = 148, 118, 99$, and $85$. Using the simple plane wave analysis discussed above as a reference guideline with $\beta = 5$ km and $W = 50$ m, we get similar values of $N$ for $\beta_2/\beta_1 = 0.56, 0.65, 0.71,$ and $0.76$. These velocity contrasts (i.e., $\beta_1 = \beta_3 = \beta_4 = 3.0$ km s$^{-1}$; $\beta_2 = 1.7, 1.9, 2.1,$ and $2.3$ km s$^{-1}$) are used in Figure 7, and comparison of the results with Figure 4a illustrates the tradeoffs between the FZ velocity contrast, FZ width, and propagation length along the fault. Although the tradeoffs are strong, they are not complete, and each parameter has its own signature. For example, the dominant period increases with increasing velocity contrast (Figure 7), increasing FZ width (Figure 4a (left)), and increasing propagation length (Figure 4a (right)). However, the overall duration of the response increases considerably with increasing velocity contrast, changes little with FZ width, and increases mildly with propagation length. Nevertheless, the results demonstrate the importance of examining jointly volumes in parameter space associated with sets of parameters before reliable conclusions on FZ properties can be made. It is also clear from the discussion above that the ratio of the velocities inside and outside the FZ, rather than individual values, is the relevant velocity parameter governing the waveforms of seismic FZ waves. Thus uncertainty in velocity of the country rock (related to typical errors of locations and origin times of earthquakes) can translate into uncertainty in the velocity of the FZ layer. Individual velocity values affect the arrival time of the phases, and these can provide important constraints for waveform inversions [Michael and Ben-Zion, 1997].

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Variable receiver depth; \( r=5\text{km}, W=50\text{m} \)

Variable receiver depth; \( r=7.5\text{km}, W=150\text{m} \)

Figure 6. Effects of variable receiver depth. Numbers on traces give receiver depth below the free surface in meters. (left) \( r=5\text{km}, W=50\text{m} \), and \( x=0.5W \). (right) \( r=7.5\text{km}, W=150\text{m} \), and \( x=2W \).
Figure 7. Effects of variable velocity contrast. Velocity of FZ layer from top to bottom is 1.7, 1.9, 2.1, and 2.3 km s\(^{-1}\). Velocity of all other media is 3 km s\(^{-1}\), \(r = 5\) km, \(W = 50\) m, \(x = 0.5W\), and \(Q_2 = 100\). Compare results with Figure 4a.

long-period oscillations. Recent summaries of observations and various theoretical frameworks can be found in, for example, Julian [1994], Chouet [1996], and Kaneshima et al. [1996]. Here we use the model of Figure 1 to calculate acoustic radiation generated by a unit slip at the interface between an extreme low-velocity layer and a bounding elastic rock. The model may provide a useful idealization for aspects of wavefields associated with narrow linear volcanic structures, such as dikes containing a mixture of magma and gas, cracks filled with fluid and gas, mid-ocean ridges, etc.

Figure 11 shows time histories and amplitude spectra of displacement seismograms in a half space with a 20 m wide vertical layer having various extreme low values of \(P\) wave velocity. The media parameters are \(\alpha_1 = \alpha_3 = \alpha_4 = 5.0\) km s\(^{-1}\), \(\alpha_2 = 0.1-0.4\) km s\(^{-1}\), \(W_2 = 20\) m, and \(Q_2 = 100\). The source and receiver coordinates are \((x_s = 0, z_s = 4.0\) km) and \((x = 5W_2, z = 0)\). The source is a simple Heaviside time function as in the previous section, and thus all interesting aspects of the response are generated solely by the assumed structure. The dike-type layer produces long duration, slowly decaying, oscillations with some features similar to those of observed volcano seismograms. However, in contrast to typical volcanic tremors that begin with low-level motion and have delayed peak amplitudes (Y. Fukao, personal communication, 1996), the simulated seismograms start with high-amplitude sharp (direct) waves. Using different quarter spaces on the two sides of the dike can lead to an emergent initial head wave and delayed peak amplitude on the slower bounding rock. However, this does not provide a realistic configuration for the present context nor a full solution for the above discrepancy, and is not pursued here. Another difference between the calculations and typical volcanic tremors is that the persisting oscillations, simulated in Figure 10 for a structure with a single dike and a simple singular source, do not propagate to distances larger than \(\sim 10-20\) times the width of the low-velocity layer (see also Figures 3, 10, and 14), whereas volcanic tremors are often recorded at much larger distances.

Figures 12 and 13 show calculations for models with a dike-type layer having a width of 100 and 200 m, respectively. The dominant periods of the model seismograms in Figures 11-13 increase with the velocity contrast and the width of the low-velocity layer, as in Figures 4 and 7. The seismograms have a spectral peak around 1 Hz and a series of weak overtones.

Figure 14a gives results generated by a model with an extreme low-velocity dike-type layer and an adjacent transition zone in an elastic half space. The media parameters are \(\alpha_1 = \alpha_4 = 5.0\) km s\(^{-1}\), \(Q_1 = Q_4 = 1000\), \(\alpha_2 = 0.1\) km s\(^{-1}\) in the top time and spectral records and 0.2 km s\(^{-1}\) in the bottom ones, \(W_2 = 20\) m, \(Q_2 = 50\), \(\alpha_3 = 1.5\) km s\(^{-1}\), \(W_3 = 100\) m, and \(Q_3 = 300\). The source and receiver coordinates are \((x_s = 0, z_s = 4.0\) km) and \((x = 10W_2, z = 0)\). The incorporation of a low-velocity transition layer adds richness to the simulated response, and the results are qualitatively reminiscent of observed volcano seismic data. Figure 14b gives results for models with wider dike-type layer having a similar extreme low-velocity \((\alpha_2 = 0.1\) km s\(^{-1}\)). In these cases the oscillations persist longer, and appreciable amplitudes can be observed further away from the low-velocity structure. As mentioned above, the calculated seismograms differ in important respects from typical volcanic tremors. Nevertheless, the simulations show prominent trapped waves which may exist in the vicinity of dikes and high-permeability fracture zones.

Observation of such waves may provide useful information on structural properties, source offset from the structure, etc.

The calculations discussed so far focus on wave guide effects associated with a simple nonrepeating source in a single dike or crack zone. More complicated, longer duration waveforms can be obtained by superposition of similar results. This is illustrated in Figure 15 where we give two examples of velocity seismograms generated by periodic spatial repetition of a dislocation source at the boundary of low-velocity dike-type structure and the surrounding rock. In the model generating Figure 15 (top), each repeating element of the low-velocity structure is a 20 m wide vertical layer with \(P\) wave velocity and quality factors of 0.2 km s\(^{-1}\) and 100, respectively. The (spatially repeating) dislocation source is at a depth of 3 km, and it is characterized, as in all previous calculations, by a unit Heaviside step function \(h(t)\). The horizontal distance between repeating members of the model
is 15 km. The receiver coordinates are \((x = 60 \text{ m}, z = 0)\). The host half space is characterized by \(P\) wave velocity and quality factors of 5.0 km s\(^{-1}\) and 1000, respectively.

In the model producing Figure 15 (bottom), the basic repeating dike-type structure has an extreme low-velocity layer with properties as before and an additional 100 m wide adjacent transition layer with \(P\) wave velocity and quality factors of 1.5 km s\(^{-1}\) and 300, respectively. Here the receiver coordinates are \((x = 120 \text{ m}, z = 0)\). The separation distance between repeating elements and other model parameters are as before. The results of Figure 15 show long sequences of sharp arrivals embedded in oscillatory wavetrains, in qualitative agreement with general features of volcanic tremors (see, for instance, examples given by Julian [1994]). The calculations may represent approximately the response of a large volcanic domain to a regional source of deformation, generating slip at multiple boundaries between dikes or cracks and the surrounding host rock. The simulations can be made more realistic by convolving the solution with nonlocalized source function having finite extent in space and time.

5. Discussion

Theoretical and observational works show that fault zones and other low-velocity structures with regular material interfaces can generate head waves in the slower velocity

**Figure 8a.** Effects of FZ attenuation coefficient. Numbers on traces give values of \(Q_2\). Propagation length is \(r = 5\) km.

**Figure 8b.** Time and spectral velocity seismograms for strong FZ attenuation. Numbers on traces give values of \(Q_2\).
medium and trapped waves in the vicinity of the low-velocity structure [e.g., McNally and McEvilly, 1977; Fukao et al., 1983; Ben-Zion and Aki, 1990; Li et al., 1990, 1994]. Proper identification of first arriving FZHWs (and their exclusion from data sets used by algorithms employing laterally homogeneous solutions) can remove biases and errors from velocity structures, earthquake locations, and fault plane solutions derived from near-fault data [Ben-Zion, 1989; Ben-Zion and Malin, 1991]. Explicit incorporation of FZHWs in travel time tomography can further improve the accuracy and stability of derived velocity structure and related quantities [Ben-Zion et al., 1992]. Waveform modeling of the various

Figure 9. Seismograms generated by sources outside the FZ in a three-media model. Numbers on traces give source offset from the FZ in units of the FZ width. Propagation length is $r = 10$ km.

Figure 10. Seismograms generated by sources outside the FZ in a four-media model. Numbers on traces give source offset from the FZ in units of the FZ width. Propagation length is $r = 10$ km.
seismic FZ phases can provide an imaging tool of FZ structure with an unmatched resolution of a few meters to a few tens of meters [e.g., Leary et al., 1987; Li and Leary, 1990; Leary and Ben-Zion, 1992; Hough et al., 1994]. For comparison the smallest structural feature resolved by the usual seismic tomography is on the order of 500 m [e.g., Eberhart-Phillips et al., 1995]. An accurate determination of FZ properties such as width and seismic velocity provides a basic foundation for various other studies. In addition, recent theoretical calculations [Andrews and Ben-Zion, 1997; Harris and Day, 1997] indicate that seismic FZ waves interacting with dynamic rupture in laterally heterogeneous structures can have important consequences for various issues of fault dynamics, including the heat flow paradox, short rise time in earthquake slip histories, strong ground motion, and more.

The numerous possible applications and dynamic effects of seismic FZ waves are coupled with strong sensitivity of the waves to a large number of parameters. Proper use of seismic FZ waves requires a clear understanding of their phenomenology and tradeoffs in the governing parameters. Toward this goal we have used the 2-D analytical solution of Ben-Zion and Aki [1990] to examine the dependency of the waves on basic geometrical and material properties. The results show that there are significant tradeoffs between propagation distance along the structure, FZ width, velocity contrast, source location within the FZ, and Q.
Figure 12. Temporal and spectral displacement seismograms for a 100 m wide vertical layer with an extreme low-velocity. Layer $P$ wave velocity from top to bottom is 0.2, 0.4, 0.75, and 1.0 km s$^{-1}$. The velocity of all other media is 5 km s$^{-1}$. The receiver is at ($x = 300$ m, $z = 0$) and $Q_2 = 100$.

calculations demonstrate that the wavefield depends strongly not only on receiver distance from the low-velocity structure but also on receiver depth below the free surface. The strong parameter tradeoffs imply that failure to account properly for some of the above effects can produce errors and scatter in the other derived parameters. This amplifies the usual need (associated with the inherent nonuniqueness of modeling) for careful quantitative analysis employing a large amount of data. As pointed out in previous works, the generation and observation of seismic FZ waves involve sources and receivers that are in the vicinity of the low-velocity structure. The simulations done here (e.g., Figures 3, 9, 10, and 14) indicate, however, that the zone connecting sources generating the waves and receiver positions with appreciable wave amplitudes can be over an order of magnitude larger than the FZ width. Thus visual inspection of the region where seismic FZ waves exist may provide some general information, but it cannot be used to conclude accurately on properties of the structure and source offset from the FZ.

Calculations for structures with an extreme low-velocity layer (Figure 11-15) show prominent long-duration trapped waves which may exist in seismic data associated with volcanic and geothermal areas. As in the context of tectonic fault zones, waveform modeling of such phases can provide high-resolution information on dike-like structures. In addition, the recognition and modeling of FZTWs in volcano seismic data may help to separate basic structural effects from those of exotic sources.

We note that the parameters governing seismic FZ waves do not form an orthogonal coordinate system in a parameter space. Changes in values of a given parameter can modify the effects of other parameters (e.g., Figures 4a and 4b). The
Figure 13. Temporal and spectral displacement seismograms for a 200 m wide vertical layer with an extreme low-velocity. Layer P wave velocity from top to bottom is 0.4, 0.75, 1.0, and 1.25 km s$^{-1}$. The velocity of host rock is 5 km s$^{-1}$. The receiver is at $(x = 600 \text{ m}, z = 0)$ and $Q_2 = 100$.

analysis done here involved mostly separate examinations of the interaction among pairs of parameters. A fuller study should consider joint variations in a volume of the entire parameter space. We have examined a simple parameter space associated with uniform plane-parallel layers. Extensions of our basic parameter space to 3-D cases of $P$, $SV$, and $SH$ waves in FZ models with irregular geometries and nonuniform material properties are given by Leary et al. [1991] and Igel et al. [1997]. Additional results for complex FZ structures are given by Leary et al. [1993], Huang et al. [1995], and Li and Vidale [1996]. The various possible complications from 3-D wavefields in irregular FZ structures not covered here strengthen further the need for careful analysis.

Igel et al. [1997] found from 3-D numerical calculations that correlated heterogeneities of FZ material properties (correlation length $\geq$ FZ width) destroy the ability of the fault system to act as a wave guide for FZTWs. In addition, horizontal surface layer with properties (width and velocity) similar to those of the FZ layer diffuses the FZTWs and inhibits the ability to obtain useful observations at the surface. On the other hand, small-scale random heterogeneities of material properties in the FZ and moderate geometrical perturbations have little effect on the amplitude and waveform of FZTWs. The simulations of Igel et al. [1997] suggest that observed FZTWs average out small FZ irregularities. It thus appears that the existence of FZTWs at given locations implies that the FZ structure is fairly uniform for the recorded range of wavelengths and hence can be modeled effectively for those wavelengths by average property, plane-layered media.

As argued a number of times in this work, a proper quantitative analysis of observed seismic FZ waves should involve the simultaneous modeling of many waveforms. An
efficient modeling procedure requires a computational scheme that can, on one hand, account for the various basic phases expected to exist near low-velocity FZ structure, while being at the same time fast enough to be used as a forward modeling kernel in a systematic inversion of a large data set. The above considerations, the calculations of the present work, and the limited previous observational studies of Leary and Ben-Zion [1992] and Hough et al. [1994] suggest that the 2-D analytical solution employed in this work may provide an appropriate basic tool for large-scale modeling of observed seismic FZ waves. The solution can be used to model portions in both $P$ and $S$ waveforms containing body waves, head waves, reflections from the free surface and FZ walls, and trapped waves. The model allows for arbitrary number of vertical layers, fine lateral FZ structure, large propagation distance, attenuation effects, variable source location inside and outside the FZ, and arbitrary receiver depth below the free surface. An important shortcoming of the solution is that it does not incorporate variations of properties along the fault. For relatively regular fault zones this can be dealt with by dividing the FZ to segments that are more-or-less uniform and treating each segment separately. In such an approach the FZ properties are modeled in a piecewise fashion, and the lengths of the various coherent segments become parameters to be determined by the data. If, on the other hand, the FZ structure is so irregular that the above approach cannot be used, it will probably not generate FZ waves at all [Igel et al., 1997].

The calculations of this work cover several combinations of basic parameters governing seismic FZ waves. It is difficult to extract from the results a conclusive quantitative summary of

![Temporal and spectral displacement seismograms](image)

**Figure 14a.** Temporal and spectral displacement seismograms for a 20 m wide dike-type layer flanked by a 100 m wide transition zone. Dike layer velocity in top and bottom seismograms and corresponding spectra is 0.1 and 0.2 km s$^{-1}$, respectively. Transition zone velocity is 1.5 km s$^{-1}$. Velocity of host rock is 5 km s$^{-1}$. The receiver is at $(x = 200 \text{ m}, z = 0)$, $Q_2 = 50$, and $Q_3 = 300$. 
Two vertical low-velocity layers; \( W_2 = 40 \), 100m, \( W_3 = 80 \), 200m

![Figure 14b. Displacement seismograms for a dike-type layer with P wave velocity 0.1 km s\(^{-1}\) and an adjacent transition zone with velocity of 1.5 km s\(^{-1}\). In top seismogram and spectrum \( W_2 = 40 \) m, \( W_3 = 80 \) m, and \( x = 200 \) m, while in bottom ones \( W_2 = 100 \) m, \( W_3 = 200 \) m, and \( x = 1000 \) m; \( Q_2 = 50 \), and \( Q_3 = 300 \).](image)

the tradeoffs between model parameters. This is because the waves exhibit strong coupled dependency on all examined material \((c_j, Q_j)\) and geometrical \((W_j, x, z, r)\) parameters, and the examined cases provide only illustrative examples rather than a complete parameter space survey. A proper quantification of the tradeoffs requires an inversion procedure which can map volumes in parameter space that have equivalent effects, within some error limits, on the waves. A study in this direction was initiated by Michael and Ben-Zion [1997], who developed a procedure for a joint inversion of P FZ head waves and S FZ trapped waves for FZ structure, locations of the earthquakes within the FZ, and distance of the seismometer from the FZ walls. The inversion is based on correlation coefficients between the observed phases and synthetic waveforms computed with the solution (1). The analysis takes initial hypocentral coordinates and properties of the bounding rocks from standard techniques, while an initial set of FZ parameters is found by forward waveform modeling. The inversion procedure then perturbs all the parameters within specified search ranges in order to maximize the correlation coefficients. Since the strong parameter tradeoffs make it difficult to assess the errors associated with the synthetic fits, the inversion employs a grid search and calculates the error function over the entire parameter space. The results can thus be used to study the structure of the error function and attach quantitative uncertainties to the best fitting parameters.

As an initial application, Michael and Ben-Zion [1997] employed the inversion procedure to model FZ trapped waves generated by a cluster of earthquakes in the Parkfield segment.
of the SAF. The events were recorded by the borehole station Middle Mountain and were located ~5 km deep and 5 km to the SE from the station. The initial model, based on previous results of Li and Leary [1990] and Leary and Ben-Zion [1992], was assumed to consist of two FZ layers between two crustal blocks. Arrival time information of various phases was used to limit the range of model parameters. The range of best fitting parameters centered around the values $\beta_1 = 2.0$ km s$^{-1}$, $Q_1 = 200$, $\beta_2 = 1.3$ km s$^{-1}$, $Q_2 = 50$, $W_2 = 110$ m, $\beta_3 = 1.4$ km s$^{-1}$, $Q_3 = 60$, $W_3 = 150$ m, $\beta_4 = 1.9$ km s$^{-1}$, and $Q_4 = 200$. The correlation coefficients computed by the inversion were, however, spread over the ranges of allowed parameters, substantiating the strong parameter tradeoffs discussed in this work. The initial results of Michael and Ben-Zion [1997] suggest that the segment of the SAF between the generating earthquakes and Middle Mountain can be characterized effectively by a single FZ layer with a width of ~250 m, S wave velocity reduction with respect to the surrounding rock of ~30%, and a $Q$ value of ~50. A more comprehensive analysis on sensitivity of model parameters and results from large-scale inversion of Parkfield FZ waveforms will be reported in a future work.

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