A reassessment of the rupture characteristics of oceanic transform earthquakes

Rachel E. Abercrombie1 and Göran Ekström
Department of Earth and Planetary Sciences, Harvard University, Cambridge, Massachusetts, USA
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[1] We investigate the long-period source spectra of oceanic transform earthquakes and find that previously proposed slow rupture components can be explained as artifacts generated by the modeling procedure. We use low-frequency (≤20 mHz) Rayleigh and Love waves to calculate the amplitude spectra of five earthquakes on the Romanche and Chain transform faults in the equatorial mid-Atlantic Ocean. We find that errors and approximations in the centroid depth, focal mechanism, and earth structure at the source have significant effects on the shape of the source spectra. If global catalog values and an average crustal model are assumed, the spectra exhibit apparent anomalous energy at long periods which has previously been interpreted as a result of slow rupture. We recalculate the source spectra using precise, independently determined depths and moment tensors and a more realistic oceanic crustal structure in the source region. The resulting source spectra are flat at long periods with no indication of anomalous long-period energy. Our results imply that oceanic transform earthquakes do not commonly have detectable slow rupture components.

INDEX TERMS: 7209 Seismology: Earthquake dynamics and mechanics; 7215 Seismology: Earthquake parameters; 7230 Seismology: Seismicity and seismotectonics; 8123 Tectonophysics: Dynamics, seismotectonics; 8150 Tectonophysics: Evolution of the Earth: Plate boundary—general (3040); KEYWORDS: Romanche transform, long-period surface waves, broadband body waves, earthquake nucleation, earthquake precursor, segmentation


1. Introduction

[2] Earthquake rupture is frequently described as a scale-invariant process [Kanamori and Anderson, 1975; Abercrombie, 1995] in which the slip and rupture velocities, and stress drop are independent of the earthquake magnitude. Hence the rupture dimension is proportional to the rupture duration, and both are observed to increase with increasing seismic moment in a simple fashion. Earthquakes that have longer durations than can be predicted from this scaling relationship have been very loosely termed "slow". They are interpreted as having either a relatively low slip velocity, or rupture velocity, or both. A wide range of observations of slow slip have been made, some associated with seismic earthquakes and some not. They range from small creep events on the San Andreas fault [e.g., Linde et al., 1996] to larger-scale slow slip on subduction interfaces (e.g., the anomalous strain observed to begin about 15 min before the great 1960 earthquake [Kanamori and Cipar, 1974; Cifuentes and Silver, 1989]) and to the recently documented slow slip lasting over 6 months beneath Tokai, Japan [Fujii et al., 2001].

[3] A number of observations have led to the inference that slow earthquakes are common, and may even represent the majority of earthquakes, on oceanic transform faults. In this study we investigate some of the observations on which this inference is based. We conclude that although some earthquakes in oceanic lithosphere (as elsewhere) appear anomalous, there is no evidence that slow, long-period rupture is common in oceanic transform earthquakes.

[4] Early work on ocean transform earthquakes suggested that they have relatively low slip velocities, and durations a few seconds longer than expected from standard scaling relationships [e.g., Kanamori and Stewart, 1976; Okal and Stewart, 1982]. More recently it has been proposed that ocean transform earthquakes often have a slow rupture component of the order of 100 s [Ihmle et al., 1993; Ihmle and Jordan, 1994; McGuire et al., 1996]. It is these proposed slow rupture components that are the focus of our study.

[5] The possibility of anomalously slow rupture is important as in many cases the slow rupture has been interpreted as preceding the normal speed rupture [e.g., Ihmle and Jordan, 1994; McGuire et al., 1996]. If this is the case, such observations of slow precursory slip could be useful for short-term prediction. They could also provide information about how the different composition of the oceanic and continental lithosphere control rupture nucleation.

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1Now at Department of Earth Sciences, Boston University, Boston, Massachusetts, USA.
Some earthquakes undoubtedly involve slow rupture that lasts a hundred seconds or so. The Nicaragua 1992 (MW 7.6) subduction earthquake is a classic example [Kanamori and Kikuchi, 1993; Ihmle, 1996]. The observations of such slow rupture components to earthquakes on oceanic transforms remain controversial, however, partly because the moment of the proposed slow component is relatively small and partly because they have not been confirmed by different research groups. For example, Ihmle et al. [1993] reported a slow precursor (~100 s) to the 1989 Macquarie Ridge earthquake. The results of a study by Park [1990] were inconclusive, and subsequent analyses by Kedar et al. [1994], Cifuentes [1995], and Velasco et al. [1995] failed to confirm the existence of a slow precursor. Abercrombie and Ekström [2001] reanalyzed the 1994 Romanche earthquake and suggested that the slow component proposed by McGuire et al. [1996] was simply an artifact of the modeling assumptions.

Modeling recorded seismograms to extract the source spectrum (or source time function) requires both input information and a number of assumptions. These include the depth and orientation of the source, and also the model earth structure in which the wave are excited, propagated and recorded. For example, Dziewonski and Gilbert [1974] reported an anomalous source spectrum for the deep 1970 Colombia earthquake, including a long-period precursory component of ~150 s duration. Russakoff et al. [1997] recently reexamined this earthquake and found that the anomalous source spectrum was caused by wave propagation effects unaccounted for by Dziewonski and Gilbert [1974], in particular the coupling effect of Earth’s rotation on low-frequency normal modes. It is possible that such mode coupling could be responsible for some of the anomalous long-period energy observed for the 1960 Chile earthquake by Cifuentes and Silver [1989].

The aim of the present study is to investigate how uncertainties in such modeling parameters as the earth structure, and the location and focal mechanism of the source affect the resolution of a slow rupture component of oceanic transform earthquakes. We initially focus on the 1994 Romanche earthquake because it is a well known earthquake for which a slow component has been proposed. We then extend our analysis to four nearby earthquakes on the Romanche and neighboring Chain transforms on the equatorial Mid-Atlantic Ridge (Figure 1 and Table 1). In Appendix A we include details of our broadband body wave models of these earthquakes. We conclude by reassessing the various observations that have been used as evidence of slow rupture in oceanic transform earthquakes over a wide range of timescales.

2. Calculation of the Long-Period Spectrum

A recorded seismogram $S(t)$ is the convolution of the source radiation, propagation and path effects and the instrument response. The response functions $I(t)$ of modern instruments are well known. The source radiation $R(t)$ is a function of the spatial orientation of the source (location, depth and focal mechanism) and the time dependence of the moment rate $M(t)$. The path effects $G(t)$ include the velocity structure, geometric spreading and attenuation. Hence the observed seismogram is written

$$S(t) = \dot{M}(t) \ast R(t) \ast G(t) \ast I(t),$$

where the asterisk denotes convolution.
Table 1. Earthquakes Considered in This Study

<table>
<thead>
<tr>
<th>Date</th>
<th>Time, UT</th>
<th>Latitude,°</th>
<th>Longitude,°</th>
<th>$M_H$</th>
<th>Depth,°</th>
<th>Strike,°</th>
<th>Dip,°</th>
<th>Rake,°</th>
<th>Transform</th>
</tr>
</thead>
<tbody>
<tr>
<td>14 March 1994</td>
<td>0430:07.60</td>
<td>−1.083</td>
<td>−23.929</td>
<td>7.0</td>
<td>17</td>
<td>82</td>
<td>75</td>
<td>179</td>
<td>Romanche</td>
</tr>
<tr>
<td>18 May 1995</td>
<td>0006:26.70</td>
<td>−0.950</td>
<td>−21.985</td>
<td>6.8</td>
<td>21</td>
<td>81</td>
<td>90</td>
<td>180</td>
<td>Romanche</td>
</tr>
<tr>
<td>26 Dec. 1992</td>
<td>1952:24.90</td>
<td>−0.564</td>
<td>−19.318</td>
<td>6.8</td>
<td>22.5</td>
<td>82</td>
<td>83</td>
<td>179</td>
<td>Romanche</td>
</tr>
<tr>
<td>28 Aug. 1992</td>
<td>1818:46.40</td>
<td>−0.965</td>
<td>−13.562</td>
<td>6.8</td>
<td>8</td>
<td>257</td>
<td>86</td>
<td>178</td>
<td>Chain</td>
</tr>
</tbody>
</table>

*Reported by NEIC.
* Determined from Harvard CMT analysis.
* Determined from broadband body wave modeling, depth is km below sea surface.

[10] Transforming to the frequency domain and rearranging, we obtain the source spectrum

$$M(\omega) = \frac{S(\omega)}{R(\omega)G(\omega)f(\omega)}.$$  (2)

To calculate the source spectrum, we calculate synthetic seismograms for each station that recorded the earthquake. We use a delta function source

$$M(t) = \delta(t)M_0,$$  (3)

where $M_0$ is the seismic moment, with the preferred location and focal mechanism, in a given model earth structure. The observed amplitude spectrum is then divided by the synthetic amplitude spectrum to obtain the source amplitude spectrum at each station. The mean earthquake source amplitude spectrum is calculated from the individual station estimates. In this analysis we are only concerned with the amplitude spectrum and neglect the phase spectrum. This is partly because the phase spectrum is harder to interpret but mainly because we are primarily interested in whether or not the earthquakes involve a slow component and not its timing.

[11] The synthetic seismograms are calculated using a hybrid method [Ekström, 2000] which allows us to account for, in an approximate fashion, the effects of excitation in a source structure that differs from the structure at the receiver. This is important since the earthquakes considered here are all in oceanic lithosphere whereas almost all the stations are on continental lithosphere. In this method, the fundamental mode surface waves are synthesized using ray theory, allowing us to excite the waves in one earth structure, propagate the waves across a heterogeneous Earth model, and receive them at a station located at the top of a different Earth structure. We do not include focusing and refraction in the ray calculation. Overtones are calculated using standard mode summation techniques, with corrections for lateral heterogeneity calculated using the path-average approximation [Woodhouse and Dziewonski, 1984].

[12] The long-period surface waves dominate the observed spectrum for several hours following the earthquake, and the hybrid method allows us to investigate and model the influence of the structure at the source, something which is not easily done with a strict mode summation approach. We can check the hybrid method by comparing the results to those obtained using standard mode summation methods. When the source structure is identical to the average Earth model (in our case, PREM), the hybrid seismograms coincide with those obtained by mode summation.

[13] The spectral ratio method that we adopt here is similar to the technique used previously by Ihmle and Jordan [1994] and McGuire et al. [1996]. They band pass the data and synthetics and calculate an average spectral amplitude at individual frequencies. The source spectrum of the 1994 Romanche earthquake calculated by McGuire et al. [1996] from vertical Rayleigh waves is shown in Figure 2.

[14] In Figure 2, note the increasing amplitude below 8 mHz that McGuire et al. [1996] interpreted as excess long-period energy resulting from a slow component to the rupture. An earthquake of this size without any slow component would have a spectrum shaped more like the dashed line, almost flat below ~8 mHz. To compare our technique to that of McGuire et al. [1996], we calculate a source spectrum using essentially the same assumptions. We use a source depth of 15 km, the Harvard CMT catalog mechanism [Dziewonski et al., 1981], and the PREM [Dziewonski and Anderson, 1981] structure at the source and receiver. Synthetic vertical Rayleigh waves are then calculated in the frequency range 1.4 to 30 mHz, using a

Figure 2. Source spectrum of the 1994 Romanche earthquake. The symbols show the spectrum of McGuire et al. [1996]. The solid curve is our Rayleigh wave spectrum, and the 95% confidence limits of the mean. The diamond is the Harvard CMT moment (4.11 $\times$ 10$^{19}$ N m), and the dashed line is the spectrum corresponding to a 40 s triangular source time function.
three-dimensional velocity model for wave propagation [Dziewonski and Woodward, 1992].

[15] We edit the seismograms using 6-hour windows, or shorter, depending on the noise level. We then apply a cosine taper and Fourier transform the records to the frequency domain. The spectra at each station are smoothed with a 1 mHz wide running window, and only the frequency range in which the signal is at least three times the noise level (determined from the 6-hour time window preceding the earthquake) is used in the subsequent analysis.

[16] We are able to reproduce the results of McGuire et al. [1996] within the 95% confidence limits, and also observe an apparent increase in source spectral amplitude at low frequency. The differences between the two spectra probably result both from the different analysis techniques and the different selection of stations at each frequency. Thirty vertical and twenty-one transverse recordings are used in our analysis, and there is significant variation from station to station. We can also calculate the spectrum from higher-frequency (10–30 mHz) body waves, but consider them subject to more uncertainty than the longer-period surface waves. They are not needed to resolve a slow rupture component and so we do not include them here.

3. Effect of Model Assumptions on the Long-Period Spectrum

[17] We consider the effect of three principal assumptions made in calculating the source spectrum from long-period surface waves: the focal mechanism, the source depth, and the model earth structure at the source. We use both vertical Rayleigh and transverse Love waves; previous studies have concentrated on the former [e.g., Ihmélé, 1994; Ihmélé and Jordan, 1994; McGuire et al., 1996].

3.1. Source Mechanism

[18] The Harvard catalog CMT best double-couple focal mechanism for the 1994 Romanche earthquake has a dip of 61° to the north. This dip is relatively shallow for an earthquake on an oceanic transform fault. Broadband modeling of eight earthquakes on the Romanche transform (including the 1994 earthquake, see Appendix A) finds an average dip of 80° to the south, with little variation [Abercrombie and Ekström, 2001]. The dip of these earthquakes may not be resolvable in the standard CMT procedure. The synthetic seismograms calculated with the preferred broadband source models fit the data used in the CMT calculations with RMS errors less than 10% larger than the minimum values obtained in the CMT inversions. To investigate the effect of source mechanism on the long-period spectrum we calculate synthetic waveforms for a pure vertical strike-slip source and use them to recalculate the source spectrum that best fits the long-period spectra. In Figure 3a we compare source spectra obtained by assuming a vertical strike-slip mechanism and the 61° dipping catalog mechanism. Since we are using the same structure at the source and the receiver the hybrid technique and standard mode summation produce identical results. Both the vertical Rayleigh and transverse Love waves are affected by the mechanism. The rate of increase in amplitude with decreasing frequency of both wave spectra is greater when the dip is decreased.

3.2. Source Depth

[19] The international agencies locating earthquakes typically fix the depth of oceanic earthquakes at 10 or 15 km because of the lack of constraining data from regional stations. In previous studies of oceanic earthquake source spectra, the depths were constrained to 15 km because it is hard to constrain the depth of shallow earthquakes from the long-period surface waves [e.g., McGuire et al., 1996]. Recent modeling of earthquakes on the Romanche transform shows that many earthquakes are significantly deeper than 15 km [Abercrombie and Ekström, 2001]; see Appendix A. We therefore calculate synthetic seismograms for two different fixed source depths and use them to determine the best fitting source spectra of the 1994 Romanche earthquake. Again, since we use the same structure at the source and receiver the hybrid method gives the same results as standard mode summation. We fix the mechanism to vertical strike slip. The shape of the source spectrum obtained from the vertical Rayleigh waves is strongly dependent on the assumed depth (Figure 3b). The spectrum corresponding to a shallower source has a steeper slope at long periods. The spectrum of a typical
earthquake of this magnitude would be flat at these frequencies; if the spectrum based on a source at 15 km depth is interpreted as the source spectrum, the anomalous long-period spectrum could be considered as evidence for a slow rupture component. The transverse component shows negligible depth dependence, as expected from the relative insensitivity of the excitation of Love waves to source depth. The spectra of Rayleigh and Love waves agree best for the deeper source. Both wave types are generated by the same earthquake and so discrepancies between them are an indication that some aspect of the modeling is incorrect.

3.3. Earth Structure at the Source

[20] Modeling of surface waves is typically performed using the same simple 1-D earth structure at the source and receiver, while frequency-dependent phase corrections are used to adjust the arrival times for propagation in a laterally heterogeneous earth. Mode summation can therefore be used to calculate the synthetic seismograms. The 1-D model most commonly used is PREM [Dziewonski and Anderson, 1981], which has a ~21 km crust below a 3 km water layer. This is the model used so far in the present study. Oceanic crust averages only about 6 km in thickness, and most oceanic transform faults are beneath water depths of more than 3 km. The PREM crustal model is therefore a poor approximation of the earth structure at the source of oceanic transform earthquakes. To investigate whether the source structure has a significant effect on the shape of the long-period source spectrum, we compare the frequency dependence of the waves excited by a source at a range of depths in two different earth models. We calculate surface wave excitation in PREM, as before, and also in an oceanic earth model based on PREM but with a 5-km water layer above a more appropriate 6-km-thick crust [Mooney et al., 1998]. The excitation functions for both vertical Rayleigh and transverse Love waves from a pure vertical strike-slip source are shown in Figure 4.

[21] For this source orientation, the frequency dependence of the excitation functions is independent of azimuth. The Love wave excitation is nearly independent of the source depth but the frequency dependence is different in the two earth models. The Rayleigh wave excitation is dependent on both the source depth and the earth structure. The excitation at one depth in the oceanic model is approximately equal to that in PREM at a depth about 5 km greater. The different depth dependence of the excitation of Rayleigh and Love waves is consistent with the source spectra shown in Figure 3b. The frequency-dependent excitation \( \mathcal{R}(\omega) \) is a term in the denominator of equation (2) and so these differences in excitation of the synthetics translate directly into the resulting source spectrum.

[22] The excitation functions are dependent on the source mechanism, and so the depth dependence observed here for strike-slip earthquakes cannot be extrapolated to other earthquakes with different mechanisms. For example, as part of their study of the 1994 Java subduction earthquake, Abercrombie et al. [2001] investigate the shape of the long-period source spectrum of a shallow dipping thrust earthquake. They find that for such a source orientation, less long-period energy is needed to model observed spectra using a dip of 12° than 7°. Also, for this source geometry, varying the depth from 15 to 20 km has a relatively small effect on the shape of the inferred source spectrum.

4. Recalculation of the Long-Period Source Spectrum

[23] The assumed focal mechanism, depth, and model earth structure are found to have a significant effect on the shape of the derived long period source spectrum. The oceanic crustal structure is preferable to the 1-D PREM, but choosing the most accurate centroid depth and mechanism is more difficult. To avoid trade-offs between source spectral shape and other parameters in the long-period modeling we use depths and mechanisms obtained independently in a recent broadband modeling study [Abercrombie and Ekström, 2001]. The high-resolution data and the good station coverage enabled us to obtain reliable parameters (see Table 1 and Appendix A). Also, the consistency of the results for neighboring earthquakes...
supports their reliability. Abercrombie and Ekström [2001] recalculated the source spectrum of the 1994 Romanche earthquake using the oceanic velocity structure, a depth of 17 km (12 km below the seafloor) and the source mechanism in Table 1. The original spectrum (PREM, catalog CMT and 15 km depth) and the recalculated one are compared in Figure 5.

The Rayleigh and Love wave spectra are more similar for the new model parameters than for the original. Also, the increase in amplitude at low frequencies in the original source spectra is not present in the recalculated ones. These recalculated spectra are well matched by the theoretical spectrum of a 25 s box-car source, close to the source duration obtained in the broadband modeling [Abercrombie and Ekström, 2001]. The uncertainties in the model assumptions appear large enough to account for the hypothesized slow component, suggesting it is simply an artifact of the modeling process.

In order to investigate whether these observations are robust we consider four additional earthquakes in the equatorial mid-Atlantic Ocean, two on the Romanche transform and two on the neighboring Chain transform. All four earthquakes have well constrained mechanisms and depths [Abercrombie and Ekström, 2001]. The two earthquakes in 1992, one on the Romanche (Mw6.8) and one on the Chain (Mw6.8) transform, were also studied by Ihmlé and Jordan [1994] and Ihmlé [1994], who identified them as having slow rupture components. The other earthquakes in the region that were included in these earlier studies are older and too poorly recorded for conclusive detailed analysis here.

The source spectra of the May 1995 (Mw6.8) Romanche earthquake are shown in Figure 6. First, we calculate the source spectra using the CMT catalog mechanism, depth and PREM. The vertical and transverse spectra are in poor agreement, and both deviate from the predicted duration of 10 s obtained from the broadband body waves. The vertical spectrum has a steep slope, suggesting a slow rupture component. When we recalculate the source spectra using the oceanic earth model and the source model obtained from the body wave modeling (21 km below sea surface, see Appendix A), these anomalies disappear. The spectra from the vertical and transverse waves agree well, and neither shows a significant deviation from the predicted source spectrum. The source spectra calculated for the 1992 earthquake on the Romanche transform using first the CMT catalog mechanism, depth, and PREM, and second the broadband model show very similar systematic differences (Figure 7).

The 16 February 1996 (Mw6.6) earthquake on the Chain transform does not have similarly steep source spectra (Figure 8). The depth from the broadband modeling (13 km below the sea surface, see Appendix A) is

Figure 5. Source spectra of the 1994 Romanche earthquake. Solid lines are the means, and dotted lines are the 95% confidence limits of the mean. (top) Original spectra and (bottom) spectra recalculated using the preferred model (Table 1). The dashed curve is for a 25 s source time function, with the moment obtained from an inversion with the source parameters fixed.

Figure 6. Inferred source spectra, as in Figure 5, for the 1995 Romanche earthquake. The broadband body wave model parameters are given in Table 1.
smaller than those of the three Romanche earthquakes. Correcting for depth, mechanism and source structure does improve the fit of the transverse and vertical wave spectra to a short-duration source, but there is not the major difference seen for the previous earthquakes (Figures 6 and 7). This observation is consistent with our hypothesis that the slope in the long-period spectrum increases as the depth is underestimated. Ihmlé [1994] found a small slow component to the 1992 earthquake (Figure 9a). The effect of the reanalysis is similar to the 1996 earthquake (Figure 8), but sufficient to remove any anomalous long-period energy. Analysis of all five earthquakes is, therefore, consistent with an incorrect model earth structure at the source combining with an underestimated depth to produce source spectra with apparent anomalous long-period energy.

5. Discussion of Earthquake Long-Period Spectra

[28] We investigate the factors affecting the frequency dependence of earthquake spectra obtained from long-period surface waves and find that apparent anomalous energy at long periods can commonly be attributed to deficiencies in the modeling procedure. The principal causes are errors in the estimates of the depth and the use of an inappropriate model of the shallow earth structure at the source. In addition to this, a number of other factors need to be considered when assessing the validity of source parameters obtained from modeling long-period surface waves.

5.1. Attenuation

[29] Attenuation alters the frequency content of seismic waves. The apparent frequency dependence of attenuation depends on the assumed attenuation as a function of depth in the earth because lower frequency waves are sensitive to greater depths. In our calculations we used the QL6 attenuation structure obtained by Durek and Ekström [1996] for the whole earth, which has been used for earthquakes in the Harvard CMT catalog starting in 1994 [Dziewonski et al., 1994]. Prior to that, the attenuation structure from PREM was used in the CMT catalog [Dziewonski et al., 1981]. We also calculate spectra using the PREM attenuation structure but the differences are insignificant.

[30] More localized oceanic attenuation structures, such as those derived for the Pacific Ocean by Katzman et al. [1998] and Chen et al. [1999] have higher Q in the upper mantle lid and low-velocity zone than the global average models and may be more appropriate for surface wave propagation within the Atlantic Ocean basin. J. J. McGuire (personal communication, 1999) has found that applying a localized Q model, determined for the Pacific Ocean, to the seismograms recorded for the 1994 Romanche earthquake results in an anomalous source spectrum regardless of source depth. It would not be appropriate to use an
most of our stations are located more than 60° oceanic attenuation structure in the present analysis since
parameters are given in Table 1. Chain earthquake. The broadband body wave model
Inferred source spectra, as Figure 5, for the 1992
Figure 9.

Figure 9. Inferred source spectra, as Figure 5, for the 1992 Chain earthquake. The broadband body wave model parameters are given in Table 1.

The similarity of the spectra of the closely located
1996 Chain earthquake show a similar pattern to those of the Romanche earthquakes. They
are not as similar to those of either of the Romanche earthquakes as the Romanche earthquakes are to one another, presumably because the earthquake is further away and the paths are not so similar.

5.2. Interstation Variability and Directivity

The 95% confidence limits plotted in Figures 2 and 5–9 are large, though similar to those of McGuire et al. [1996] and Ihmlé and Jordan [1994], and they limit our resolution of the long period spectrum. The wide confidence limits imply a large scatter between the recordings at individual stations. This could be caused partly by source directivity; for example, the 1994 Romanche earthquake propagated unilaterally to the east [Abercrombie and
Ekström, 2001; McGuire et al., 1996]. However, the other
two best recorded earthquakes considered (1995 and 1996) did not have strong directivity but their confidence limits reflect similar interstation variability. Directivity such as that observed for the 1994 Romanche earthquake must contribute to the uncertainties but it cannot be the principal
cause.

To investigate the uncertainties resulting from the azimuthal distribution, we select nine stations, three each from three different azimuths: northeast (Europe), southeast (Africa, Australia, etc.), and northwest (North America). We plot the source spectra calculated from these stations using the preferred source models (broadband body wave depth and oceanic structure) in Figure 10 for the 1994 and 1995 Romanche transform earthquakes.

The similarities between the individual spectra for the two Romanche earthquakes (Figures 10a and 10b) are striking, especially for the vertical Rayleigh waves. The stations to the northeast have spectra increasing in amplitude with frequency, whereas those recorded in the north-west show spectra decreasing with increasing frequency. The same is observed to a lesser extent for the transverse component. The southeast stations are similar to the northeast ones, but with larger scatter, probably because they are typically at greater epicentral distances.

The 1996 Chain earthquake show a similar pattern to those of the Romanche earthquakes. They are not as similar to those of either of the Romanche earthquakes as the Romanche earthquakes are to one another, presumably because the earthquake is further away and the paths are not so similar.

The similarity of the spectra of the closely located
1994 and 1995 Romanche earthquakes enables us to con-
sider the 1995 earthquake as an empirical Green function
for the larger 1994 earthquake. In standard empirical Green function analysis, the spectrum of the large earthquake $S_1(\omega)$ is divided by that of a colocated smaller event $S_2(\omega)$, and corrected for the scalar moment of the smaller event $M_2(0)$, to obtain an estimate of the source spectrum of the larger earthquake,

$$\hat{M}_{1,2}(\omega) = \frac{S_1(\omega)}{S_2(\omega)} M_2(0).$$

We do not simply perform an empirical Green function analysis because a slow component in the smaller earthquake might have removed the one from the larger. We thus divide our preferred source spectra of the 1994 earthquake, by those of the 1995 earthquake, after correction for model earth structure, etc., to obtain an improved
estimate of the source spectrum of the larger (1994) earthquake,

\[
\dot{M}_{1,2}(\omega) = \frac{M_1(\omega)}{M_2(\omega)}M_2(0). \tag{5}
\]

The result (Figure 10c) shows that the interstation variability is very similar for both earthquakes, especially below 12 mHz. At higher frequency the distance separating the two earthquakes is probably more significant. Any directivity present in the 1994 but not the 1995 earthquake spectra will also be more prominent.

[37] The effectiveness of this division in reducing the interstation variability for the two earthquakes implies that structural effects not included in the models are the cause of this scatter. Lateral velocity heterogeneity causes significant refraction and focusing of long-period surface waves [e.g., Woodhouse and Wong, 1986]. In addition, lateral variations in \( Q \) will also have systematic effects on the amplitudes. Neither of these two effects are modeled in our synthetic waveforms. The relative magnitude of these variations again significantly limits the resolution of long-period part of the earthquake source spectrum.

5.3. Implications for the Harvard CMT Catalog

[38] The Harvard CMT catalog is now commonly used by many researchers, and so we need to consider the implications of our study on the reliability of this catalog. Using the PREM crustal structure for oceanic transform earthquakes

**Figure 10.** Plots showing the interstation variability. The black solid spectra are from stations to the northwest (HRV, SCZ, and SJG), the black dashed spectra are from stations to the northeast (PAB, TAM, and ECH), and the gray spectra are from stations to the southeast (PAF, CRZF, and NWAO).
will, in general, cause the body waves to underestimate the true depth and the mantle waves to overestimate it. Constraining the depth at 15 km, as is commonly done, results in errors in the orientation and nondouble-couple component of the focal mechanism as the inversion attempts to find a “best” fit. The difference between the catalog depths and mechanisms and those determined by Abercrombie and Ekström [2001] is relatively small, however, and indicates the level of uncertainty in the global catalog. The uncertainties are largest where the assumptions used in the catalog analysis are least accurate. Ekström [1987] and Patton [1998] have noted relatively large uncertainties for earthquakes in areas of thick continental crust. Here we find related problems in areas where the crust is significantly thinner than in PREM.

6. Are Oceanic Transform Earthquakes Slow?

[39] Our results show that a slow rupture component of a few hundred seconds is not required to match the observed long-period spectra of oceanic transform earthquakes as has previously been proposed. Thus it is likely that oceanic transform earthquakes may not as commonly involve such slow rupture as has been claimed. The term “slow earthquake” does not only refer to such events, however, but to phenomena over a wide range of timescales that are not necessarily related. Therefore we briefly reassess the published evidence for slow rupture in oceanic earthquakes. We classify reported slowness by increasing timescale. We begin by considering the earliest reports by Kanamori and Stewart [1976] and then continue to increasing timescales.

6.1. Slow Slip or Rupture Velocity?

[40] Some oceanic transform earthquakes have been termed “slow” because they have durations of only a few seconds longer than average. For example, Kanamori and Stewart [1976] identified two earthquakes on the Gibbs Fracture Zone (northem Mid-Atlantic Ridge) as having low $m_b$ (measured at 1 s) with respect to $M_S$ (measured at 20 s). One interpretation of this observation is that the earthquakes radiated more energy than expected at periods of 20 s and so could be considered slow. Okal and Stewart [1982] also presented examples of oceanic earthquakes with low $m_b$ relative to $M_S$, possibly for similar reasons. Kanamori and Stewart [1976] assumed a trapezoidal source time function, and calculated relatively long rise times ($\sim 10$ s), and low slip velocity ($\sim 0.2$ m/s) which support their interpretation of slow rupture. The rupture velocity, however, was found to be normal. Since it is now thought that such a trapezoidal source time function is an oversimplification, it is not clear how reliable these estimates of slip velocity and risetime are.

[41] Kaverina et al. [1996] defined the “creepex” of an individual earthquake as the deviation of $M_S - m_b$ for that earthquake from the global average. They performed a global study and found that strike-slip earthquakes tend to have higher $M_S$ relative to $m_b$ than dip-slip earthquakes, consistent with the larger number of nodal $P$ arrivals of the former at teleseismic distances. In addition, Kaverina et al. [1996] found that creepex was positive for mid-ocean ridge and fracture zone earthquakes, principally in the Southern Ocean, and negative for subduction zones. It remains unclear, however, how much of this difference is a source effect and how much is a result of different propagation paths. In addition, the strike-slip earthquakes on the Mid-Atlantic Ridge show no significant difference from the global average value.

[42] Sheaver [1994] considered the difference between radiation at 20 s ($M_S$) and longer periods ($\geq 60$ s) in a global study focusing on surface waves. He also found a tendency for oceanic transform earthquakes to have higher levels of long-period radiation relative to $M_S$ than other earthquakes. Again, the majority of these earthquakes were in the Southern Ocean where network coverage and detection are poorest.

[43] Ekström and Dziewonski [1988] investigated global relationships between $M_S$ and $M_b$, and found that oceanic earthquakes had lower $M_S$ relative to $M_b$ than subduction zone earthquakes. They were unable to determine from their analysis whether the various trends were the result of source processes, or simply regional differences in the excitation and propagation of 20 s surface waves.

[44] Recently, the ratio of radiated seismic energy ($E_S$) to seismic moment has been used to investigate the ratio of high- to low-frequency source radiation. In a global study, Choy and Boatwright [1995] found that oceanic strike-slip earthquakes have some of the highest ratios of $E_S$ to $M_b$. This would suggest that oceanic strike-slip earthquakes are relatively “fast” or high stress drop. Choy and McGarr [2002] note that high $E_S$ to $M_b$ earthquakes may be occurring in relatively strong oceanic lithosphere. Pérez-Campos et al. [2001] noted that although some oceanic strike-slip earthquakes have high ratios of $E_S$ to $M_b$, some also have low values. These studies would seem to imply that there are both anomalously slow, and fast, earthquakes on oceanic transforms, as there are in many other tectonic environments.

[45] A particularly unusual earthquake, with anomalously long seismograms containing excess long-period energy, occurred two weeks after the great 1960 Chile earthquake, at the southern end of the rupture zone [Kanamori and Stewart, 1979]. Kanamori and Stewart [1979] interpreted the earthquake as a series of slow events triggered by a moderate-sized earthquake, all occurring on the subducted extension of the Chile Rise transform fault system. The data are too limited to constrain the nature of this complex earthquake well, however, and no similar events have been described since. This earthquake demonstrates that anomalous slip can occur on transform faults, at least after subduction.

[46] The various observations summarized above are intriguing but do not present a consistent pattern of anomalously slow faulting behavior on oceanic transform faults. Further work is required to establish which of these observations are robust and can be confidently be attributed to anomalous earthquake properties.

6.2. Prince Edward Island Earthquake

[47] McGuire and Jordan [2000] described the 28 April 1997 Prince Edward Island earthquake ($M_b$6.8) as an analogue of the Romanche earthquakes. The relatively high noise level and low quality of the broadband
recordings of this Southern Hemisphere earthquake mean that we cannot resolve a precise mechanism or depth. The long-period spectrum is relatively poorly resolved and does not include significant excess long-period energy [McGuire and Jordan, 2000]. McGuire and Jordan [2000] also noted that the Prince Edward Island earthquake began with a small-amplitude onset, starting at the high-frequency origin time and lasting about 15 s. Many other earthquakes begin with small amplitude onsets [e.g., Abercrombie and Mori, 1994; Ellsworth and Beroza, 1995]. McGuire and Jordan [2000] used regional broadband onsets to suggest that the small beginning of the Prince Edward earthquake is lower frequency and hence smoother than a typical small earthquake. This contrasts with the interpretation of the Romanche 1994 earthquake by McGuire et al. [1996] in which the proposed slow rupture preceded the high-frequency onset by over 100 s. McGuire et al. [1996] stacked low-pass-filtered records of the 1994 Romanche earthquake, to show the slow precursor, but the signal-to-noise ratio is poor, and we do not find these observations convincing.

[45] McGuire and Jordan [2000] also proposed that the Prince Edward Island earthquake involved a large segment jump, and liken it to two earthquakes (1971 and 1994) on the Romanche transform. Abercrombie and Ekström [2001] were able to fit the Romanche 1994 earthquake seismograms with no such jump (and could not fit the seismograms including a jump), and concluded that it was merely an artifact of mispicking nodal P arrivals, see Appendix A. Forsyth et al. [1986] noted a similar segment jump during the 1971 Romanche earthquake. They realized the inherent ambiguity in picking potentially nodal arrivals from sub-events in a complex rupture without waveform modeling, and so did not publish this initial interpretation (D. W. Forsyth, personal communication, 1999). We, therefore, do not believe that the Prince Edward Island earthquake provides convincing evidence for either anomalously slow rupture or a segment jump.

### 6.3. Aseismic Slip and Seismic Coupling

[49] The term "slow" earthquake has also been used to describe slip events recorded on strain meters and by GPS, with durations of minutes (e.g., in Italy [Crescentini et al., 1999]), to hours and days (e.g., on the San Andreas fault in California [Linde et al., 1996] and in Japan [Kawasaki et al., 1995]). Slow creep has also been detected on subduction interfaces lasting for months, e.g., in Cascadia [Dragert et al., 2001] and on the Tokai subduction zone [Fujii et al., 2001]. These events are completely aseismic, although some appear to affect nearby seismicity patterns, and involve no slip at the speeds of typical earthquakes. Both slip velocity and rupture velocity are inferred to be very slow.

[50] As yet, observations of such aseismic events on oceanic faults have not been possible. It is probable that some aseismic slip occurs on oceanic transform faults, however, because the summed seismic moment is significantly less than plate motions would predict. Abercrombie and Ekström [2001] found the seismic slip since 1920 to be about 50% of the plate motion on the Romanche and Chain transforms. The length of the seismic catalog is always a concern in such comparisons as large earthquakes might have occurred earlier. Similar deficits have been found on other transform faults [e.g., Bergman and Solomon, 1988; Wolfe et al., 1993; Boettcher and Jordan, 2001], however, and it is less likely that no large earthquakes occurred anywhere on the entire global oceanic transform system.

### 7. Conclusions

[51] We investigate the effects of uncertainties in the modeling of the shape of the long-period source spectrum of five oceanic transform earthquakes. Slow rupture components have been reported previously for four of these earthquakes. We find that the derived source spectra are significantly affected by the choice of centroid depth, source geometry and the earth model in the source region. Global catalog values and average crustal models are inadequate to resolve small anomalies in the long-period spectra of oceanic transform earthquakes. When more precise, independently determined, model parameters are used, the spectra of the five earthquakes studied do not exhibit anomalous energy at long periods. We conclude that the previously reported long-period slow (and sometimes precursory) rupture components of oceanic transform earthquakes are simply artifacts of uncertainties in the modeling procedure. Our review of other reports of anomalous oceanic earthquakes suggests that some of them may involve slow rupture, but that events of this type may not be as common as has previously been proposed.

#### Appendix A: Broadband Body Wave Modeling of Earthquakes: Did the 1994 Romanche Earthquake Jump Fault Segments?

[52] Abercrombie and Ekström [2001] modeled the body waves of 14 earthquakes on the Romanche and Chain transforms. Their model of the 1994 Romanche earthquake does not include the 80 km segment jump proposed by McGuire et al. [1996]. The models of the long-period spectra presented in this paper depend on the source parameters determined from this broadband body wave modeling and so we include figures showing our resolution (Figures A1–A6). We show the results for modeling the 1994 Romanche earthquake. For comparison we also show the results of modeling two of the other earthquakes considered in the present study, 1995 Romanche and 1996 Chain earthquakes. We also show the results of our attempts to model the 1994 earthquake with the jump in location proposed by McGuire et al. [1996].

[53] We model teleseismically recorded broadband body waves to obtain improved depths and moment tensors. For each earthquake we use all available recordings of P and SH waves at Global Seismic Network (GSN) and Geoscope stations. We correct for the instrument responses to obtain displacement seismograms filtered between 1 and 150 s. We then invert for the moment tensor, source time function, depth and directivity using the method developed by Ekström [1989]. We use $f^*$ of 0.6 s for P waves and 3.0 s for S waves. We use a layered velocity model over a half-space with a 6-km crust and 5 km of water. We fix the sources at the NEIC epicenters. We do not include the CMT solution as a constraint on the inversion, even though that is
Figure A1. Preferred, single fault, model of the 1994 Romanche earthquake at selected stations. The observed (solid lines) and synthetic (dashed lines) seismograms are shown for each station. The amplitudes of the seismograms ($\mu$m) are given at the top right. The arrows mark the $P$ onsets, and the vertical lines indicate the length of seismogram used in the inversion. Directivity is shown by the arrow on the $P$ wave mechanism. The source time function is shown at the bottom.
Figure A2. Model of the 1994 Romanche earthquake with two offset subevents. We use the same data in the inversion as for our preferred model. We add a second subevent at the offset location and time determined by McGuire et al. [1996]. The layout is the same as Figure A1. The mechanisms and source time functions of the two subevents are shown at the bottom. The relative locations of the two subevents are also shown. The first subevent is close to that from our preferred model and the inversion puts negligible moment into the second subevent.
Figure A3. Model of the 1994 Romanche earthquake with two offset subevents. This model is essentially the same as Figure A2, except that we shorten the duration of the first subevent to force the inversion to put moment onto the second fault. The layout is the same as in Figure A2. Note the poor fits to the P wave seismograms in the SE (LBTB, BOSA, and SUR) as well as the poorer fits to the SW (NNA and LPAZ).
an option. We initially allow nondouble-couple components, but these are so small that we restrict the inversions to double-couple solutions.

[54] Figure A1 shows our preferred model for the 1994 Romanche earthquake. Our initial $P$ picks agree with those of McGuire et al. [1996]. A single fault model, involving an initial small subevent (their subevent A) followed by two larger ones, fits the principal features of the waveforms well at all azimuths. The high rupture velocity is below the $S$ wave speed in oceanic mantle, although it implies strong unilateral rupture to the east. McGuire et al. [1996] interpret the first large arrivals (negative in the NW (e.g., HRV) and SE (e.g., LBTB), and positive in the SW (e.g., LPAZ) and NE (e.g., TAM)) as $P$ phases from the first large subevent (their subevent B) and locate it 80 km NE of the onset. Such a large segment jump has never previously been observed.

[55] In order to investigate whether the model of McGuire et al. [1996] would fit the data better than our single fault model, we perform inversions including a second subevent on a fault with the onset time and location determined by McGuire et al. [1996]. The inversion result (Figure A2) has negligible slip on the second fault, and its orientation is not well constrained. The slip on the first fault is almost identical to that in our preferred model (Figure A1). We then tried another series of inversions in which we shortened the duration of possible slip on the first fault, to force the large subevents onto the second fault. As shown in Figure A3, this model results in significantly poorer fits (the variance doubles). In particular note that the inversion cannot match the stations to the SE and does a relatively poor job to the SW. Further analysis shows that there is no focal mechanism with the polarities and relative amplitudes of the subevent B identified by McGuire et al. [1996]. Why a single fault solution works so well is demonstrated in Figure A4. This focal mechanism has clear $P$ arrivals to the NW and NE, as observed. The stations to the SE are nodal, and the first large negative arrival is $sP$. To the SW the $P$ arrival is relatively small, and the large positive arrival is $sP$. It seems likely that the phases interpreted by McGuire et al. [1996] as the onset of subevent B to the SE and SW actually correspond to reflected phases from the first large subevent. The time delays between $P$ and $sP$ thus give the appearance of a segment jump.

[56] McGuire et al. [2002] note that this modeling could be affected by seafloor topography along the transform faults. Wiens [1987] showed that a dipping seafloor could produce reflected arrivals not predicted by horizontal layered models (such as the one used here) and these could be mistaken for source complexity. The fact that we obtain simple source models for these earthquakes suggests that this is not a serious problem.

[57] Figures A5 and A6 show the 1995 Romanche and 1996 Chain earthquakes, respectively. The 1995 earthquake has a focal mechanism almost identical to the 1994 Romanche earthquake, and the waveforms look like a single subevent from the 1994 earthquake. McGuire et al. [2002] also found that the source orientations of these two earthquakes are very similar. The seismograms from the 1996 Chain earthquake look very different, corresponding to a shallower depth and a slightly different focal mechanism. Comparing these two earthquakes demonstrates the good resolution of the body wave modeling in depth and mechanism for these well-recorded earthquakes.
Figure A5. Model of the 1995 Romanche earthquake. Note the similarity in mechanism to the 1994 earthquake and how the seismograms correspond to those from each individual subevent of the 1994 earthquake. Layout as in Figure A1.
Figure A6. Model of the 1996 Chain earthquake. Note the significant differences in the seismograms, compared with the Romanche earthquakes, despite the small change in location, depth, and mechanism, demonstrating the resolution of the modeling technique.
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References


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R. E. Abercrombie, Department of Earth Sciences, Boston University, 685 Commonwealth Avenue, Boston, MA 02215, USA. (rea@bu.edu)

G. Ekström, Department of Earth and Planetary Sciences, Harvard University, 20 Oxford Street, Cambridge, MA 02138, USA.