Source scaling of earthquakes in the Shumagin Region, Alaska: time-domain inversions of regional waveforms

Geoffrey A. Abers,* Xiaoxing Hu† and Lynn R. Sykes†

Lamont-Doherty Earth Observatory of Columbia University, Palisades, NY 10964, USA

Accepted 1995 March 23. Received 1995 February 19; in original form 1994 June 15

SUMMARY

The scaling of pulse duration to seismic moment is estimated for earthquakes along an interplate thrust zone, from digital waveforms recorded by short-period and broad-band instruments of the East Aleutian (Shumagin) Seismic Network. We measure pulse duration using an empirical Green's function technique based on damped time-domain deconvolution. From several thousand events, 22 earthquakes with magnitudes 3.0–7.0 and depths 23–56 km are found to give reliable estimates of pulse duration. Durations are also determined directly from one-parameter non-linear inversions, for a variety of simple functional forms of source time functions. Symmetric source pulses (boxcar or triangle shapes) fit waveforms better than an asymmetric model \[ t \exp \left( -2t/D \right) \] for most (62 per cent) of the waveform pairs, while the asymmetric model fits best for only 8 per cent of the data. Pulse duration increases with the size of events, from 0.1 to 10 s over the seismic moment \( M_0 \) range of \( 10^{14} \) to \( 3 \times 10^{19} \) Nm. When normalized by the cube root of seismic moment, pulse durations show \( -8 \times \) variation; comparable static stress drop estimates range from 0.2 to 135 MPa. Contrary to predictions of some laboratory and theoretical studies, earthquakes at the deepest part of the thrust zone do not show significantly higher stress drops than do shallower events. Rupture properties, however, show a strong dependence on earthquake size. The three largest events \( (M_0 > 5 \times 10^{18} \) Nm) have the three longest normalized durations, on average 3.8 times longer than those for smaller events. The durations require smaller events to have \( 10-100 \times \) larger static stress drops, or \( -4 \times \) faster rupture velocities, or some combination of the two. Possibly, the largest events rupture both strong and weak patches while smaller events just rupture strong patches on the fault surface. The characteristic dimension that separates large from small events, 3–15 km, is comparable to characteristic wavelengths of Pacific basin bathymetry and may reflect the influence of the subducted sea-floor upon fault-zone heterogeneity.

Key words: Alaska, faulting, Green's functions, source time functions, subduction.

INTRODUCTION

It is often argued that earthquake source scaling relationships are independent of event size for most recorded earthquakes. Seismic moment \( M_0 \) varies with the cube of radius of fault rupture \( (r^3) \) for many recorded earthquakes (e.g. Kanamori & Anderson 1975; Scholz 1982), although the relationship may break down for the largest and smallest events (Scholz 1982; Fletcher et al. 1986; Dysart, Snoke & Sacks 1988). At intermediate sizes, i.e. magnitudes between 3 and 7.5, it is often argued that \( M_0/r^3 \) is roughly constant for a given tectonic setting, as expected if slip varies linearly with fault diameter (Kanamori & Anderson 1975; Scholz 1982). Because of the way global seismic measurements differ from regional observations, however, it is rare that consistent measures of source dimension are determined simultaneously for both smaller (magnitude <5) and larger (magnitude >6) earthquakes. We present here analysis of a consistent data set that includes both types of earthquakes, with magnitudes varying

* Now at: Department of Geology, University of Kansas, Lawrence, KS 66045, USA.
† Also at: Department of Geological Sciences, Columbia University, New York.

© 1995 RAS
from 3.0 to 7.0 along a single subduction zone segment. These data show characteristic differences between $M_d/r^3$ estimates for larger and smaller events.

Except for those earthquakes where aftershock zones or ground breakage are well defined, estimates of source size are derived from source pulses seen on seismic waveforms, which are interpreted with simple source models. Traditionally, these estimates have been made by measuring spectral corner frequencies (e.g. Brune 1970). Recent studies have shown that source corner frequencies are difficult to separate from propagation effects for small earthquakes (e.g. Anderson 1986), leading to large uncertainties and biases in static stress drop estimates. As an alternative, several workers have applied empirical Green's function (EGF) techniques to correct for propagation effects (e.g. Mueller 1985; Frankel et al. 1986; Hough et al. 1991). These techniques derive source time functions from waveforms by deconvolving the waveform of a nearby small event of the same source mechanism and relatively short pulse duration, thus removing path and receiver effects with a minimum number of assumptions about the source. EGF methods form the basis of the present study.

SEISMICITY OF THE SHUMAGIN REGION

The Shumagin Region (Fig. 1) is off the south-eastern end of the Alaska Peninsula, where the Pacific plate subducts beneath North America. Plate convergence here is normal to the arc at 65 mm yr$^{-1}$ (DeMets et al. 1990). The main interplate thrust zone, the region where earthquakes show thrust motion and occur at the plate boundary, extends from 25 to 45 km depth (Reyners & Coles 1982). Seismicity extends to 250 km depth, but events below 45 km depth show intraplate deformation rather than thrusting (Abers 1992a). Between the early 1960s and 1993 five events with magnitudes greater than 6 occurred in the region, all since 1983. All of these earthquakes show thrust faulting and are consistent with slip on the plate interface. Although some historical evidence exists for great earthquakes in previous centuries (Davies et al. 1981), the largest recorded by seismographs was a $M_w = 7.4$ earthquake in 1917 (Estabrook & Boyd 1992).

The Shumagin Islands are also one of the few places globally where on-land seismometers have been placed directly above an interplate thrust zone, and have provided local observations of earthquake rupture at subduction zones (Figs 1 and 2). House & Boatwright (1980) fit strong motion accelerograph records at Sand Point (SAN in Fig. 1) with a quasi-dynamic source model for two moderate-sized earthquakes. The events ($m_b = 5.8$ and 6.0; crosses in Fig. 1) occurred at depths between 37 and 43 km, at the deep end of the thrust zone. From these data and teleseismic short-period waveforms, they suggested that the deep part of the thrust zone in the Shumagin gap is characterized by

Figure 1. Eastern Aleutian (Shumagin) seismic network (EASN), Alaska. Triangles denote seismic stations. Except for SAN, which had an intermediate-period/broad-band sensor, all other stations consist of short-period, three- or vertical-component instruments. Crosses signify EASN catalogue locations for 1974 earthquakes studied by House & Boatwright (1980) and Mori (1983). Other symbols indicate earthquakes studied in this paper, each corresponding to a different Event Group as described in Table 1. 1985 sequence, hexagons; 1991 sequence, circles; 1993 sequence, stars; other events on thrust surface, squares; intraplate events within subducting plate, inverted triangles. Hypocentres are relocated by Abers (1992a) and Abers et al. (1995). Grey hexagons and circles show additional well-located aftershocks of 1985 and 1991 sequences, respectively.
events with stress drops near \(-100\,\text{MPa}\). Taber & Beavan (1986) calculated dynamic stress drops for two moderate events in a 1983 earthquake sequence (\(M = 6.3\) and \(5.6\)) at depths of \(26\,\text{km}\), and found comparatively low dynamic stress drops (<\(10\,\text{MPa}\)). Both studies suggested that stress increases with depth in the main thrust zone, consistent with dependence of friction on normal stress predicted from laboratory measurements (e.g. Scholz, Molnar & Johnson 1972). All of these results are based on only four magnitude 5.8–6.3 events, and other interpretations are possible. Mori (1983) recalculated dynamic stress drops for moderate-sized events throughout the eastern Aleutians, using teleseismic waveforms. He argued that the high stress drops of House & Boatwright (1980) were a consequence of the low rupture velocities they obtained from teleseismic seismograms (0.55 times the shear-wave velocity), and he preferred values of \(10–20\,\text{MPa}\). Overall, Mori observed a wide scatter in dynamic stress drops (0.5–50 MPa) for thrust zone events. Considerably more data now exist, particularly from smaller events recorded by a digital seismograph network established by the Lamont-Doherty Earth Observatory (Davies et al. 1981; Abers 1992a, 1994), as well as improved analytical techniques. These data are examined here.

**METHODS**

An empirical Green’s function (EGF) technique is used to determine pulse durations from digitally recorded \(P\) waveforms. A small event is chosen that has a similar hypocentre and focal mechanism to the event of interest, and is treated as a Green’s function for propagation from the source. After selection of a suitable pair of events (a ‘large’ target event and a smaller EGF event), pulse durations are estimated in two ways (in this context, the descriptor ‘large’ should not be taken as an indication of absolute size). First, a time-domain deconvolution technique is used (Appendix A) to remove the signal of the EGF event from that of the large event, and the pulse duration \((D)\) of the larger event is estimated from its deconvolved source pulse. At an early stage in the present study, experiments showed that time-domain techniques gave more robust results than frequency-domain analysis for the low dynamic range, band-limited waveforms used here (Abers 1992b). Sipkin & Lerner-Lam (1992) have shown that frequency-domain deconvolution can lead to biases in quantities derived from deconvolved waveforms, in part because the filters used to stabilize deconvolutions remove significant information at low frequencies. Also, small differences between the EGF and the true Green’s function can cause time-dependent errors, which are more easily handled by time-domain than frequency-domain stabilization methods.

For a given earthquake, the relationship between the time series for the seismogram of a large event at station \(k\), \(h_k\), and that for the corresponding EGF seismogram, \(g_k\), can be written as a discrete linear convolution:

\[
h'_k = \sum_{j=0}^{n_k} g'_j s'_j,
\]

where \(h'_k\) is the \(i\)th datum of the large-event seismogram \((i = 1, 2, \ldots, n_k)\), \(g'_j\) is the \(i\)th point of the corresponding EGF, and \(s'_j\) is the \(j\)th point of the source pulse \(s_k\) \((j = 1, 2, \ldots, m)\). In other words, the source pulse \(s_k\) is assumed to be that time series which, when convolved with the EGF waveform, recovers the seismogram for the large event \(h_k\). This system of equations can be expressed in matrix form, as

\[
h = Gs,
\]

where \(G\) is defined in the Appendix. If it is reasonable to assume that \(s\) is the same for a suite of seismograms, then the system of eqs (2) can be used to simultaneously invert all seismogram pairs for a single source pulse \(s\). For uncorrelated and Gaussian data errors, standard inverse methods (e.g. Tarantola & Valette 1982) give a solution for

**Figure 2.** Vertical cross-section through Shumagin region showing seismic stations (triangles) and distribution of seismicity (open circles) within 100 km of section. Hypocentres were relocated by Abers (1992a) and Abers et al. (1995). Solid circles denote events studied in this paper. Aftershocks of the 1991 May 30 earthquake are omitted because their depths are poorly constrained. The earthquake indicated by arrow, event 6, has a strike-slip focal mechanism and is probably within the downgoing plate; event 16 is nearby. The location of section A–A’ is shown in Fig. 1. Vertical and horizontal scales are equal.
a single seismogram pair of the form

$$
\mathbf{s} = (\mathbf{G}^\top \mathbf{W} + \mathbf{G}^\top \mathbf{l}_m)^{-1} \mathbf{G}^\top \mathbf{w}. \tag{3}
$$

Here $\mathbf{l}_m$ is an $m \times m$ identity matrix, $\mathbf{W} = \sigma^2 \mathbf{l}_m$ is the inverse covariance matrix of the data, $\sigma^2$ and is the observed power in the $k$th seismogram of the large event. We have parametrized the damping terms by $\beta$, the moment ratio of the large event to that of the EGF event, and $\varepsilon$, a damping factor (see Appendix A for details). The error scaling factors ($\sigma$, $\varepsilon$ and $\beta$) are an appropriate parametrization if multiplicative uncertainties in $\mathbf{G}$ dominate. The solution is controlled by one free parameter, $\varepsilon$, a damping term that limits the importance of small eigenvalues of $\mathbf{G}$ and which controls the trade-off between resolution and accuracy (larger $\varepsilon$ results in greater accuracy but lower resolution). Different values of $\varepsilon$ are explored in all of our solutions.

A second approach, referred to as direct inversion, is to assume a simple functional form for $\mathbf{s}$ and directly solve (2) for parameters that define that function (Appendix B). These inversions avoid most of the problems involved in deconvoluting methods, by drastically reducing the number of free parameters to one or two and by directly solving for the desired quantity (pulse duration $D$). In the solutions attempted here, $\mathbf{s}$ is defined by a scalar amplitude $A$ and a duration $D$, so that (2) becomes

$$
\mathbf{h} = A \mathbf{G}_m(D) = A \mathbf{f}(D) \tag{4}
$$

where $\mathbf{m}$ gives the normalized shape of $\mathbf{s}$ and $\mathbf{f}(D) = \mathbf{G}_m(D)$.

Eq. (4) is solved to determine a probability density function (pdf) for $D$ using the fully non-linear inverse methods of Tarantola & Valette (1982). The dependence of the solution $A$ is removed by analytic integration (Appendix B), and then the marginal pdf for $D$ is calculated by a grid search. Three functional forms are tested: a boxcar pulse (instantaneous rise and decay), a ramp or triangular pulse (linear rise followed by linear decay at same rate), and a ‘Brune’ pulse ($t \exp(-2t/D)$) resembling the far-field pulse of Brune (1970).

**DATA**

**Instrumentation**

We use digital $P$ waveforms recorded by the East Aleutian (Shumagin) Seismic Network (EASN) between 1982 July and 1991 June for well-located earthquakes near the interplate thrust zone (Figs 1 and 2). During that period the EASN consisted of 15 to 18 remote telemetered stations (5 to 10 after 1990 August) and one central station at Sand Point (SAN in Fig. 1), with few known changes to instrumentation. All stations had either high-gain, three-component or vertical-component, short-period (SP) seismometers (Geospace HS-10, 1 Hz natural frequency) or force-balanced accelerometers. After 1990, the station SAN also included a high-gain, three-component Guralp CMG-4, broad-band (BB) sensor with a 0.05 Hz corner. Before 1989, SAN included three low-gain intermediate-period (IP) channels (Fig. 3). Signals were digitized at 100 samples per second by a 12-bit digitizer, for an effective dynamic range of 96 dB for SAN and 40–70 dB for the telemetered stations. The range of instrumentation permitted the digital recording

![Figure 3. Amplitude response to displacement (in counts mm⁻¹) as function of frequency for typical short-period seismograph (SAN-SP), the Sand Point intermediate-period system (IP) and the broad-band system (BB). The broad-band system replaced IP in 1990.](https://academic.oup.com/gji/article-abstract/123/1/41/571199)
be on scale on the same instrument, the EGF events had to be within 0.6–2 magnitude units of those of the large events. Even though more than 1000 potential event pairs pass the selection rules, only about 25 per cent of them has useful seismograms for both large events and EGF events, and only 20 large events gave stable results. As a result we focus on large- and moderate-sized earthquakes with magnitudes larger than 3.5, and rely heavily on the IP/BB instruments.

Event pairs are required to have similar focal mechanisms, and the same first motion polarities at all stations. We also require visual similarity of waveforms between the large event and EGF events, and we require both waveforms to be recorded by the same sensor. Formal similarity conditions are not enforced (Aster & Scott 1993) because stable results are sometimes obtained from events with complex sources but low similarity. The final determination depends on the results of deconvolution and direct inversion; only stable results are retained (Table 1).

P waves were studied rather than S waves because they are likely to be less influenced by propagation and multiple phase effects. P-wave window lengths varied with event size and sensor type, from 3–6 s for events of magnitude 5.0–5.9 recorded on IP instruments to 1.5 s for SP waveforms of events of magnitude 3.0–3.9. For events 1 and 2, the entire P wave train was used up to 1–2 s before the S onset. For some records that were clipped on vertical component, the waveforms on the horizontal components were used.

### 1993 Events

As this project neared completion, a sequence of events occurred in 1993 May \((M_w = 6.9; \text{Abers et al.} 1995)\) directly beneath the Shumagin Islands, including the largest recorded event in the Shumagin Islands since 1948. The network was no longer operating and only strong motion accelerographs recorded these events locally. The accelerometers did not record sufficient long-period energy for any potential EGF events to be useful, so durations could not be estimated in the same way as the other events. However, because these seismograms contain substantial information about the rupture behaviour of large events, pulse durations estimated from teleseismic waveform modelling are used to include them in the study. These durations are estimated using standard inversion techniques of broadband body waves (Abers et al. 1995; Estabrook, Jacob & Sykes 1994).

### Table 1. Event parameters.

<table>
<thead>
<tr>
<th>Event ID</th>
<th>Date/Time (HR,MM)</th>
<th>Latitude (°N)</th>
<th>Longitude (°E)</th>
<th>Depth (km)</th>
<th>Magnitude</th>
<th>Event Type</th>
<th>Group</th>
<th>Mechanism</th>
</tr>
</thead>
<tbody>
<tr>
<td>01</td>
<td>05/30/91 13:17</td>
<td>54.21</td>
<td>161.46</td>
<td>32.0</td>
<td>M_SSC</td>
<td>6.7</td>
<td>89</td>
<td>2</td>
</tr>
<tr>
<td>02</td>
<td>10/09/85 09:33</td>
<td>54.68</td>
<td>159.44</td>
<td>29.0</td>
<td>M_SSC</td>
<td>6.5</td>
<td>110</td>
<td>1</td>
</tr>
<tr>
<td>03</td>
<td>10/26/85 15:59</td>
<td>54.74</td>
<td>159.43</td>
<td>27.6</td>
<td>M_SSC</td>
<td>5.5</td>
<td>99</td>
<td>1</td>
</tr>
<tr>
<td>04</td>
<td>11/14/85 22:17</td>
<td>54.65</td>
<td>159.56</td>
<td>23.0</td>
<td>M_SSC</td>
<td>5.4</td>
<td>92</td>
<td>1</td>
</tr>
<tr>
<td>05</td>
<td>06/01/91 08:56</td>
<td>54.29</td>
<td>161.41</td>
<td>52.3</td>
<td>M_PDE</td>
<td>5.3</td>
<td>80</td>
<td>2</td>
</tr>
<tr>
<td>06</td>
<td>05/02/87 19:21</td>
<td>54.78</td>
<td>159.92</td>
<td>37.6</td>
<td>M_SSC</td>
<td>5.0</td>
<td>49</td>
<td>5</td>
</tr>
<tr>
<td>07</td>
<td>10/09/85 14:16</td>
<td>54.77</td>
<td>159.45</td>
<td>26.6</td>
<td>M_PDE</td>
<td>5.0</td>
<td>9</td>
<td>1</td>
</tr>
<tr>
<td>08</td>
<td>05/30/91 15:44</td>
<td>54.24</td>
<td>161.40</td>
<td>23.3</td>
<td>M_PDE</td>
<td>4.9</td>
<td>42</td>
<td>2</td>
</tr>
<tr>
<td>09</td>
<td>10/10/85 22:53</td>
<td>54.81</td>
<td>159.58</td>
<td>27.4</td>
<td>M_SSC</td>
<td>4.9</td>
<td>59</td>
<td>1</td>
</tr>
<tr>
<td>10</td>
<td>09/25/85 16:29</td>
<td>54.69</td>
<td>159.79</td>
<td>31.4</td>
<td>M_SSC</td>
<td>4.8</td>
<td>30</td>
<td>1</td>
</tr>
<tr>
<td>11</td>
<td>08/18/82 06:32</td>
<td>54.97</td>
<td>160.17</td>
<td>42.1</td>
<td>M_PDE</td>
<td>4.8</td>
<td>11</td>
<td>4</td>
</tr>
<tr>
<td>12</td>
<td>11/15/85 15:58</td>
<td>54.69</td>
<td>159.66</td>
<td>30.6</td>
<td>M_SSC</td>
<td>4.6</td>
<td>5</td>
<td>1</td>
</tr>
<tr>
<td>13</td>
<td>05/30/91 17:32</td>
<td>54.19</td>
<td>161.41</td>
<td>50.7</td>
<td>M_PDE</td>
<td>4.1</td>
<td>2</td>
<td>2</td>
</tr>
<tr>
<td>14</td>
<td>05/30/91 23:55</td>
<td>54.30</td>
<td>161.41</td>
<td>4.6</td>
<td>M_PDE</td>
<td>4.1</td>
<td>3</td>
<td>2</td>
</tr>
<tr>
<td>15</td>
<td>01/18/89 08:02</td>
<td>54.95</td>
<td>160.18</td>
<td>43.8</td>
<td>M_SHM</td>
<td>3.6</td>
<td>2</td>
<td>4</td>
</tr>
<tr>
<td>16</td>
<td>05/12/87 06:40</td>
<td>54.78</td>
<td>159.93</td>
<td>38.9</td>
<td>M_SHM</td>
<td>3.6</td>
<td>4</td>
<td>5</td>
</tr>
<tr>
<td>17</td>
<td>08/17/82 02:15</td>
<td>54.97</td>
<td>160.16</td>
<td>44.4</td>
<td>M_SHM</td>
<td>3.6</td>
<td>8</td>
<td>4</td>
</tr>
<tr>
<td>18</td>
<td>01/18/89 14:17</td>
<td>54.97</td>
<td>160.17</td>
<td>43.3</td>
<td>M_SHM</td>
<td>3.4</td>
<td>1</td>
<td>4</td>
</tr>
<tr>
<td>19</td>
<td>01/21/87 00:30</td>
<td>54.96</td>
<td>160.18</td>
<td>43.2</td>
<td>M_SHM</td>
<td>3.2</td>
<td>9</td>
<td>4</td>
</tr>
<tr>
<td>20</td>
<td>06/22/91 18:50</td>
<td>54.75</td>
<td>160.55</td>
<td>33.5</td>
<td>M_SHM</td>
<td>3.0</td>
<td>2</td>
<td>4</td>
</tr>
<tr>
<td>21</td>
<td>05/13/93 11:59</td>
<td>54.81</td>
<td>160.26</td>
<td>36.0</td>
<td>M_PDE</td>
<td>6.8</td>
<td>45</td>
<td>3</td>
</tr>
<tr>
<td>22</td>
<td>05/25/93 23:16</td>
<td>55.03</td>
<td>160.36</td>
<td>42.8</td>
<td>M_PDE</td>
<td>6.2</td>
<td>150</td>
<td>3</td>
</tr>
</tbody>
</table>

Unless noted, the locations and depths of events are from 3-D inversion (Abers 1992a); the fault plane solutions are from first motions using data from East Aleutian seismic network.

1. Number of stations used in magnitude calculation.
2. Event sequences: (1) 1985 sequence; (2) 1991 sequence; (3) 1993 sequence; (4) other events on thrust surface; (5) intraplate events within subducting plate.
3. From teleseismic waveform modelling (T. Boyd, 1992, personal communication)
4. From teleseismic waveform modelling (Abers, unpublished results)
5. From Harvard Centroid-Moment Tensor (CMT) Catalog (e.g. Dziewonski et al. 1981).

© 1995 RAS, **GJI** 123, 41–58
DURATION MEASUREMENTS AND TESTS

From the deconvolution results, duration (Table 2) is estimated by eye as the time between the two minima on either side of the source pulse (Figs 4–6). The errors are about 0.02–0.03 s for an impulsive pulse shape, consistent with the variance observed between stations. For the direct inversions, the duration \( D \) (Table 3) is calculated directly. The formal uncertainty is taken to be the 1-sigma half-width of the probability density function. The scalar misfit, \( \Phi \), is the residual variance normalized to the data:

\[
\Phi = \frac{(h - G_s)^T W (h - G_s)}{h^T W h}.
\]  

We only report results with \( \Phi < 0.70 \), as larger \( \Phi \) values represent a poor fit. For damped deconvolutions, \( \Phi \) is generally less than 0.1 even for \( E = 10 \), and is usually less than 0.01 for \( E = 1 \). The \( \Phi \) values are reported in Tables 2 and 3.

Table 2. Estimated moments and source durations from deconvolution.

<table>
<thead>
<tr>
<th>Event</th>
<th>( M_0 ) ( (Nm) )</th>
<th>( M_0 ) ( \text{relative \ error} )</th>
<th>( N_{ef} ) ( (N_{st}) )</th>
<th>BB/IP</th>
<th>( \Phi_{\text{MAX}} )</th>
<th>Short-period</th>
<th>( \Phi_{\text{MAX}} )</th>
</tr>
</thead>
<tbody>
<tr>
<td>01(^1)</td>
<td>3.1*10(^{19})</td>
<td>1 (1)</td>
<td>8.00-10.0</td>
<td>0.12</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>02(^2)</td>
<td>8.7*10(^{11})</td>
<td>2 (2)</td>
<td>&gt;6</td>
<td>see text</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>03(^2)</td>
<td>1.3*10(^{17})</td>
<td>2 (2)</td>
<td>0.38-0.45</td>
<td>0.20</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>04(^3)</td>
<td>2.0*10(^{16})</td>
<td>1 (1)</td>
<td>0.40-0.44</td>
<td>0.04</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>05</td>
<td>5.2*10(^{17})</td>
<td>1.16</td>
<td>0.58-0.60</td>
<td>0.11</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>06(^4)</td>
<td>9.8*10(^{16})</td>
<td>2 (2)</td>
<td>0.48-0.59</td>
<td>0.06</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>07</td>
<td>1.1*10(^{16})</td>
<td>1.15</td>
<td>0.40</td>
<td>0.08</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>08</td>
<td>3.2*10(^{16})</td>
<td>1.16</td>
<td>0.25-0.31</td>
<td>0.10</td>
<td>0.25</td>
<td>0.08</td>
<td></td>
</tr>
<tr>
<td>09</td>
<td>3.6*10(^{16})</td>
<td>1.26</td>
<td>0.37-0.50</td>
<td>0.18</td>
<td>0.37-0.43</td>
<td>0.04</td>
<td></td>
</tr>
<tr>
<td>10</td>
<td>3.5*10(^{16})</td>
<td>1.17</td>
<td>0.16-0.29</td>
<td>0.11</td>
<td>0.22-0.30</td>
<td>0.05</td>
<td></td>
</tr>
<tr>
<td>11</td>
<td>1.4*10(^{16})</td>
<td>1.12</td>
<td>0.19-0.34</td>
<td>0.03</td>
<td>0.18</td>
<td>0.18(^6)</td>
<td></td>
</tr>
<tr>
<td>12</td>
<td>2.2*10(^{16})</td>
<td>1.21</td>
<td>0.20-0.29</td>
<td>0.11</td>
<td>0.18-0.19</td>
<td>0.37(^6)</td>
<td></td>
</tr>
<tr>
<td>13</td>
<td>6.7*10(^{15})</td>
<td>1.25</td>
<td>0.23-0.25</td>
<td>0.03</td>
<td>0.22-0.25</td>
<td>0.10</td>
<td></td>
</tr>
<tr>
<td>14</td>
<td>7.2*10(^{15})</td>
<td>1.11</td>
<td>0.20-0.26</td>
<td>0.02</td>
<td>0.22</td>
<td>0.03</td>
<td></td>
</tr>
<tr>
<td>15</td>
<td>8.9*10(^{15})</td>
<td>1.19</td>
<td>0.10-0.13</td>
<td>0.01</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>16</td>
<td>4.1*10(^{15})</td>
<td>1.19</td>
<td>0.11-0.12</td>
<td>0.46</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>17</td>
<td>2.2*10(^{15})</td>
<td>1.17</td>
<td>0.15-0.17</td>
<td>0.18</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>18</td>
<td>1.0*10(^{15})</td>
<td>1.21</td>
<td>0.12</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>19</td>
<td>8.2*10(^{14})</td>
<td>1.08</td>
<td>0.18-0.19</td>
<td>0.01</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>20</td>
<td>1.2*10(^{14})</td>
<td>1.06</td>
<td>( \leq 0.10)</td>
<td>0.40(^6)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>21(^2)</td>
<td>2.5*10(^{19})</td>
<td>10-15(^5)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>22(^2)</td>
<td>1.5*10(^{18})</td>
<td>1.2-2.5(^5)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Moments calculated from spectra of local waveforms, except where CMT solutions available. Error estimates are 2-\( \sigma \) uncertainties from regressions of log (amplitude) versus frequency, so are multiplicative errors in \( M_0 \). \( N_{ef} \) is the number of Green's function events, \( N_{st} \) is the number of stations used for each result. \( \Phi_{\text{MAX}} \) is the largest value of \( \Phi \) for large \( e \) among all event-EGF pairs; in all cases small \( e \) led to \( \Phi \) less than 0.01. Ranges of durations are from different estimation parameters as described in text.

1 Broad-band/intermediate period data.
2 CMT moment.
3 Event was a double event. Moment used here is that of first subevent.
4 Measured duration of this event is at resolution limit; duration maybe be shorter.
5 Durations estimated from teleseismic waveform inversion for source time function.
6 Misfit and duration from joint inversion of all waveforms for single source pulse.

The results of the direct inversion are sensitive to the start times of the waveforms for both the large and EGF events, so we include in the inversion a parameter that allows for a small amount of uncertainty between onset times of the large and EGF events (Appendix B). Combined picking tolerances are estimated for each waveform pair, and are typically 0.1 s for IP/BB seismograms and 0.05 s for SP seismograms.

Figure 4 shows an example of results obtained from event 4 recorded on an IP seismogram. The moment ratio is 12 between the large and EGF events. The duration of the large event, obtained from the apparent width of the pulse (Fig. 4a), is 0.38-0.45 s corresponding to the different damping values, \( \epsilon \). The predicted waveforms determined from direct inversion using three source models are comparable to the observed waveform of the large event (Fig. 4b). The durations calculated from box (I), ramp (II) and Brune (III) models are 0.28, 0.32 and 0.16 s with normalized misfits 0.23, 0.20 and 0.26, respectively. The uncertainties in duration range from 0.03 to 0.07 s for the three models (Fig. 4d). The duration from the Brune-pulse...
Figure 4. Example of estimation of pulse duration using the empirical Green's function technique. (a) Result from deconvolution. Width of pulse is taken to be equal to the pulse duration $D$. Solid and dotted lines correspond to the damping factor $\varepsilon = 10$ and 1 respectively. Parts (b), (c) and (d) show direct inversion results assuming simple models. (b) Observed waveform of large-event compared with synthetic waveforms determined from inversion, for boxcar (I), ramp (II) and Brune ($t \exp (-2t/D)$) (III) pulse shapes respectively, convolved with EGF record in (c). (c) Waveform of EGF (Empirical Green Function) event. (d) Probability density function for half-duration of three types of pulse shapes. Amplitudes of vertical axes are unitless for (a) and (d); in counts for (b) and (c).

($t \exp (-2t/D)$) model is apparently shorter than those from the other two models and has the highest misfit value (Table 3). The slow decay of the Brune pulse requires short rise times ($D/2$) in order to match the long-period duration of the observed pulses.

To understand the effect of damping on resolution, we deconvolved an event with itself, i.e. using the same seismogram as both the large and EGF events (Fig. 5). When the damping value is nearly zero ($\varepsilon = 10^{-4}$), the

© 1995 RAS, GJI 123, 41–58

Figure 5. Duration resolution test by using same event as both large event and EGF event: (a) short-period data; (b) intermediate period data. When $\varepsilon = 10^{-4}$, or near-zero, the pulse should be a delta function, and the width of the spike is sampling interval. Pulse width increases with damping, demonstrating finite resolution widths. (c) A real source pulse after deconvolution with different damping values. Damping values of $\varepsilon = 1$ and 10 are used in this paper.
Figure 6. Deconvolution results of event 9 for three values of the damping parameter, ε. For asymmetric pulses and the smallest value of ε, noise can lead to systematic underestimates of actual width of pulse (Sipkin & Lerner-Lam 1992).

deconvolved pulse resembles a delta function, and the width of the pulse becomes wider as ε increases. The shortest durations that can be resolved are 0.05 s corresponding to ε of 1 and 0.10 s corresponding to ε of 10 for short-period data, and 0.08 and 0.14 s corresponding to ε of 1 and 10 for broad-band data, respectively (Fig. 5). The shortest resolvable durations are near 0.1 s. When ε is large, the accuracy of the duration estimate is poor because much high-frequency information is lost. Conversely, when ε is small, high-frequency noise obscures the pulse shape (e.g. Fig. 5c). For an asymmetric pulse shape (Fig. 6), when ε is very small, noise can lead to systematic underestimates of the width of the pulse (also see Sipkin & Lerner-Lam 1992). To avoid this bias, we found ε = 1 to 10 are a reasonable compromise between avoiding overly rough pulses and having high values of Φ.

These measurements assume that the duration of the EGF event can be ignored, which may bias results because a finite duration for EGF events can produce underestimates of the duration of the large event. Some specific event pairs may be grossly in error, but on average these biases should be small. If the EGF has a source pulse s₂ and the deconvolution results gives s_{est}, then the true pulse for the large event should be s₁ = s₂ * s_{est}, where * is the convolution operator. A typical EGF event has a moment 10–20× smaller than the large event, which (for M₀/r³ constant) should have a duration 35–45 per cent of the large event, i.e. s₂ should have a duration that is ~40 per cent of s₁. At worst this implies that the estimate s_{est} should be only 60 per cent of the correct duration, and the centroid duration of s_{est} can be shown to be 90 per cent of the full duration in this circumstance (by assuming a Gaussian form for the source pulses). Even if duration estimates are systematically low by 10–40 per cent, the significant conclusions of this work (that normalized durations are several times smaller for small events than large events) should not change. The problem is
Table 3. Duration estimates from direct inversions.

<table>
<thead>
<tr>
<th>Event</th>
<th>Box Pulse</th>
<th>D from BB/IP</th>
<th>Ramp Pulse</th>
<th>D from SP</th>
</tr>
</thead>
<tbody>
<tr>
<td>ID</td>
<td>D, s</td>
<td>D, s</td>
<td>D, s</td>
<td>D, s</td>
</tr>
<tr>
<td></td>
<td>@</td>
<td>@</td>
<td>@</td>
<td>@</td>
</tr>
<tr>
<td>01</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
</tr>
<tr>
<td>02</td>
<td>4.40-7.12</td>
<td>0.70</td>
<td>--</td>
<td>--</td>
</tr>
<tr>
<td>03</td>
<td>0.28-0.30</td>
<td>0.26 .03</td>
<td>0.32 .20</td>
<td>0.07</td>
</tr>
<tr>
<td>04</td>
<td>0.51</td>
<td>0.24 .06</td>
<td>0.40 .47</td>
<td>0.06</td>
</tr>
<tr>
<td>05</td>
<td>0.50</td>
<td>0.58 .04</td>
<td>0.50-0.62</td>
<td>0.52 .02</td>
</tr>
<tr>
<td>06</td>
<td>0.44-0.48</td>
<td>0.45 .03</td>
<td>0.52 .37</td>
<td>0.07</td>
</tr>
<tr>
<td>07</td>
<td>0.20</td>
<td>0.40 .03</td>
<td>0.24 .39</td>
<td>0.03</td>
</tr>
<tr>
<td>08</td>
<td>0.20</td>
<td>0.29 .03</td>
<td>0.26-0.32</td>
<td>0.08 .02</td>
</tr>
<tr>
<td>09</td>
<td>--</td>
<td>0.46</td>
<td>0.66 .05</td>
<td>0.22-0.34</td>
</tr>
<tr>
<td>10</td>
<td>0.00</td>
<td>0.20 .04</td>
<td>0.20-0.04</td>
<td>0.21 .44</td>
</tr>
<tr>
<td>11</td>
<td>0.08</td>
<td>0.04 .04</td>
<td>0.08 .04</td>
<td>0.17 .24</td>
</tr>
<tr>
<td>12</td>
<td>--</td>
<td>--</td>
<td>0.16 .45</td>
<td>0.01</td>
</tr>
<tr>
<td>13</td>
<td>0.12</td>
<td>0.42 .03</td>
<td>0.12 .44</td>
<td>0.08-0.11</td>
</tr>
<tr>
<td>14</td>
<td>0.12-0.23</td>
<td>0.50 .02</td>
<td>0.16-0.31</td>
<td>0.35 .01</td>
</tr>
<tr>
<td>15</td>
<td>0.06</td>
<td>0.30 .01</td>
<td>0.05</td>
<td>0.19 .01</td>
</tr>
<tr>
<td>16</td>
<td>0.10</td>
<td>0.39 .00</td>
<td>0.12-0.17</td>
<td>0.37 .01</td>
</tr>
<tr>
<td>17</td>
<td>0.08-0.13</td>
<td>0.28 .01</td>
<td>0.16 .30</td>
<td>0.02</td>
</tr>
<tr>
<td>18</td>
<td>0.11</td>
<td>0.52 .01</td>
<td>0.14</td>
<td>0.57 .01</td>
</tr>
<tr>
<td>19</td>
<td>0.09</td>
<td>0.08 .01</td>
<td>0.17</td>
<td>0.07 .02</td>
</tr>
<tr>
<td>20</td>
<td>0.07</td>
<td>0.38 .01</td>
<td>0.09</td>
<td>0.49 .02</td>
</tr>
</tbody>
</table>

D is pulse duration, with formal uncertainty \( \sigma_D \). \( \Phi \) is variance of fit residual, normalized to variance of seismogram. Dashes indicate poor results, with misfit values >0.70. Brackets indicate special event pairs; see text.

MAGNITUDES AND SEISMIC MOMENTS

Seismic moments were estimated for all of the events studied (Table 2). Where available, teleseismic moments determined in the Harvard Centroid Moment Tensor (CMT) catalogue (e.g. Dziewonski, Chou & Woodhouse 1981) were used (events 1–3, 6, 21 and 22). Event 4 was found to be a double event by teleseismic body-wave inversions (T. Boyd, private communication, 1992) and durations and moment were only determined for the first subevent.

For the remaining events, seismic moments were determined from P-wave spectra of local waveforms. The network magnitude \( M_{HUM} \) appears to underestimate earthquake size, and upward biases of \( m_b \) for small events (Ringdal 1976) make it difficult to rely on \( m_b \). P-wave amplitude spectra were determined by the multi-taper method (Park, Lindberg & Vernon 1987) from 2–5 s windows containing the dominant P pulse train. These spectra were fitted to three-parameter spectral models that include instrument response, frequency-independent attenuation (e.g. Hough et al. 1991), and corner frequency, to determine low-frequency ground displacements \( U_0 \). The logarithm of spectra, \( \log(U_0) \), are fitted in a least-squares sense between the sample frequency and 10 Hz and were accepted if \( U_0 \) estimates varied by less than a factor of four over the range of tested corner frequencies. For most events the seismic moments were determined from the IP or BB instruments, although short-period estimates were included for events 11–20.

Seismic moment \( M_0 \) is determined from each estimate of \( U_0 \) using the standard relation for P waves (Aki & Richards 1980). A smoothed 1-D velocity model (Abers 1994) is used to estimate velocities, ray angles, and geometrical spreading terms; densities are assigned by a standard velocity–density relation. The radiation pattern correction is determined from the focal mechanisms in Table 1, where given, or otherwise from a nominal interplate thrust solution (strike = 240°, rake = 90°, dip = 10–40° following the dip of the thrust zone in Fig. 2). Two-sigma formal errors in \( \log(U_0) \) are treated as multiplicative uncertainties for \( M_0 \). The final moment estimate is determined by averaging estimates from several seismograms, and uncertainties are appropriately propagated (Table 2).

Individual moment estimates vary by factors of two or three or less for a given event. Estimates are within 50 per cent of the CMT moments for two of the three events small
Figure 7. Comparison of seismic moments determined from spectral levels with catalogue magnitudes. Open circles show magnitudes calculated from East Aleutian Seismic Network (EASN) catalogue, and solid diamonds show teleseismic $m_b$ (Preliminary Determination of Epicentres or International Seismological Centre catalogues). Magnitudes and sources are from Table 1. Solid line is $M_0-m_b$ relation of Nuttli (1985).

RESULTS

Pulse durations from deconvolution

Pulse durations $D$ are given for a range of damping values $\varepsilon$, ranging from 1 to 10 (Table 2). One-third of the events have estimates made from both SP and IP/BB waveforms. The other events are either too large to be recorded on scale by the SP instruments or too small for the EGFs to be seen on the IP/BB records. When a single value is listed in Table 2 instead of a range, the duration estimates were within 0.01 s for both values of damping. Overall, SP and IP/BB data give comparable results.

Pulse durations from direct inversion

Most (77 per cent) events have misfits $\Phi$ less than 0.50 from two-parameter inversions (Table 3). These solutions explain more than half the signal variance, and their pulse durations are taken to be reliable. The formal uncertainties in pulse duration for 75 per cent events are less than 0.03 s. Using IP data, the durations of a few events (10 and 11 in Table 3) are close to the 0.04 s IP sample rate from both box and ramp models, yet have misfit values of 0.04 (i.e. the model fits perfectly). The IP data do not carry much high-frequency information, so the waveforms of large and EGF events are indistinguishable. Stable durations are usually obtained from SP waveforms.

The results from the box and ramp models are comparable. Durations from deconvolution are somewhat higher than those from direct inversions for all models, particularly for small durations where resolution problems affect the deconvolution (Fig. 8). The durations calculated for the Brune pulse $t \exp (-2t/D)$ are systematically shorter than those from the boxcar and triangle models, and have generally higher $\Phi$. The $\Phi$ for the Brune pulses are higher than the symmetric pulses for 62 per cent of the events, and lower for only 8 per cent. Hence, we only show the results of the box and ramp inversions in Table 3. Apparently, the $t \exp (-2t/D)$ model is less effective in...
fitting the pulse shape than the symmetrical pulse models. This suggests that healing pulses are comparable in duration to rise times so that far-field source pulses appear more symmetric.

**Scaling parameters**

Although traditional estimates of static stress drop (e.g. Brune 1970) provide a means to examine scaling of pulse duration to seismic moment, they depend strongly on unknown characteristics of the earthquake processes such as rupture velocity. We consider a normalized duration $D_{N} = D / D_{REF}(M_0)$, which gives the duration relative to $D_{REF}(M_0)$, the duration expected from the corresponding moment for a standard fault model. Variations in $D_{N}$ indicate the extent to which ruptures are relatively fast or slow, and are linear with respect to measurement errors in $D$. Although $D_{REF}$ depends on stress drop, it is only sensitive to the cube root of both stress drop and radius. Without correction for moderate uncertainties in these parameters have a small effect on $D_{N}$.

In the results shown in Fig. 9, $D_{REF}$ is calculated from the radius of a circular fault using the relation

$$D_{REF}(M_0) = 2(r_{REF}(M_0)/v)(1 - \sin \theta v / \alpha_v).$$

Here, $r_{REF}(M_0)$ is the reference fault radius, $v$ is a rupture velocity, $\alpha_v$ is the source-region $P$ velocity, and $\theta$ is the angle the outgoing ray makes with the fault normal. This relationship is used in many EGF studies (e.g. Mori & Frankel 1990; Xie et al. 1991) and, although it is only valid for some cases of simple circular rupture (Boatwright 1984), provides a scaling of pulse duration with earthquake size that is consistent with other studies. The fault radius is calculated by assuming the fault is a uniform stress-drop circular crack (Keilis-Borok 1957):

$$r_{REF}(M_0) = \left[ \frac{7 M_0}{16 \Delta \sigma_0} \right]^{1/3},$$

where $\Delta \sigma_0$ is a reference static stress drop. In practice, $\alpha_v$ is assumed to be 0.9 times the shear-wave velocity at the source, $\beta$, and Poisson’s ratio is assumed to be 0.25. The term $(1 - \sin \theta v / \alpha_v)$ is approximated by its spherically averaged value of 0.63, giving $D_{REF} = 1.40 \alpha_v \Delta \sigma_0^{-1}$. We assume $\Delta \sigma_0 = 3$ MPa, a typical value for interplate earthquakes (Kanamori & Anderson 1975), and $\beta = 4.22$ km s$^{-1}$, its value at 35 km depth in the 1-D velocity structure (Abers 1994). Static stress drops $\Delta \sigma$ are calculated directly from $D_{N}$, as $\Delta \sigma = \Delta \sigma_0 D_{N}^{-3}$.

**Specific events**

Event 1, the mainshock of the 1991 May sequence ($M_0 = 6.7$), occurred in the south-western part of the Shumagin region 10 months after the nearby part of the network had been decommissioned. Only waveforms from the Shumagin Island stations, 150–200 km distant, were recorded. About 15 s of the broad-band $P$-wave seismogram is on scale on the vertical component. Because the waveform is long and the waveforms are very complicated, the direct inversions did not give reliable results ($\Phi > 0.80$). The depths of this event and its aftershocks (Group 2 in Table 1) are poorly controlled by the Shumagin network, but teleseismic waveform modelling of the mainshock gave a depth of 32 ± 4 km (not shown).

Event 2 is the mainshock ($M_0 = 6.5$) of the 1985 October sequence near the Shumagin Islands. The event is close to the network, and only 8–9 s of horizontal-component seismograms are on scale from the SAN-IP instruments. Durations range from 6.3 to 8 s, reaching the window lengths, and probably give a minimum value of $D$. The calculated duration could also be determined from the box pulse inversion for three EGF events, which gave $D$ of 4.5–7.1 s with $\Phi$ of 0.58–0.67. The boxcar model fits best at the beginning of the waveform. Teleseismic waveform inversion gives a duration near 7 s and a depth of 29 km (T. Boyd, 1992, private communication); the CMT solution gives a similar depth (32 km). From these considerations we infer a duration of $7 \pm 1$ s, consistent with both the deconvolution results and the teleseismic inversions. Many events with seismic moments less than $10^{18}$ Nm are aftershocks of events 1 and 2 (Fig. 1, Table 1).

Using eq. (7) and the pulse durations obtained from deconvolution, the fault radii of events 1 and 2 are 23–29 km and 18–24 km respectively. We compare these estimates to the area of the aftershock zone as outlined by aftershocks within 10 days of the mainshocks, relocated in the 3-D velocity model (Fig. 1). Although poorly constrained by the sparse network then operating, event 1 has an aftershock zone with a radius of 25–30 km, nearly identical to that estimated from pulse durations. Event 2, directly beneath the network, has an elliptically shaped aftershock zone with dimensions 35–40 km by 25–30 km, giving an area roughly 1.5 times larger than predicted from pulse durations. The discrepancy may reflect additional energy buried in the $S$ wave train, aftershocks outside the mainshock rupture zone, or limitations of eq. (7). Nevertheless, the aftershock zones are close in size to the predictions made from pulse duration, and independently support the fault radius–pulse duration relationship (eq. 7) and the methods used to determine duration.

**DISCUSSION**

**Variation in duration with seismic moment**

Duration generally increases with seismic moment (Fig. 9). The scaling between the two, described as $D_{N}$ or $\Delta \sigma$, shows a significant difference between the largest earthquakes ($M_0 > 5 \times 10^{18}$ Nm) and smaller events. The larger events exhibit uniformly larger $D_{N}$ (or smaller $\Delta \sigma$). There are no overlaps between values for the two populations, which on average show $D_{N}$ four times larger for the three largest events ($\Delta \sigma$ is on average 50 times smaller for these events). The three largest events (1, 2 and 21, with $M_0 = 0.9–3.1 \times 10^{19}$ Nm) have the three longest $D_{N}$ of 1.4–2.2 ($\Delta \sigma = 0.3–1.0$ MPa). The smaller events, with $M_0 = 10^{14}–10^{18}$ Nm, show $D_{N} = 0.2–1.2$ ($\Delta \sigma = 2–110$ MPa). Direct inversions give similar results (Table 3). Hence, a characteristic difference exists between the scaling behaviour for larger and smaller earthquakes. The transition occurs at durations of 1–5 s, corresponding to fault radii of 3–15 km. This transition in scaling behaviour is distinct from that observed in other studies near $10^{21}$ Nm, which may separate.
Figure 9. (A) Seismic moment $M_o$ as a function of pulse duration $D$, from deconvolution results (Table 2). (B) $M_o$ as a function of normalized duration $D_N$. Solid lines and pluses denote results from IP or BB waveforms recorded at SAN. Dashed lines and $\times$ denotes results from SP data. Stars denote results without deconvolution from teleseismic waveform inversion for 1993 events. Numbers following each line indicate event numbers in Table 1 and 2. Arrow symbol indicates a maximum duration that approaches resolution limits. Dashed line labelled 'EE' is empirical $M_o-D$ correlation of Ekström & Engdahl (1989) from broad-band inversion of teleseismic waveforms.
earthquake (thick solid line) is characterized by a fault length \( r_l \) and a smaller length \( r_r \), while a small earthquake (thick stippled line) is characterized by a smaller length \( r_l \) and a high stress drop \( \Delta \sigma_2 \). Both nucleate where \( \tau_s \) is near \( \tau_f \), so often \( \Delta \sigma_1 < \Delta \sigma_2 \).

One explanation for the relatively low stress drops of larger events is that they rupture both strong and weak portions of the fault, while smaller events rupture strong portions. One might expect rupture to nucleate where shear stresses on the fault are high, close to the frictional strength of the fault (Fig. 10). Small earthquakes only break the region near the nucleation point. Rupture of larger earthquakes continues into nearby regions, which either are conditionally stable and cannot nucleate earthquakes (e.g. Scholz 1990) or have shear stresses below the static frictional stress of the fault segment until the rupture reaches those regions (Fig. 10). In either case, much of the rupture zone of large earthquakes could have low shear stresses. This description resembles the asperities of Lay & Kanamori (1981), although it is not clear that areas of high stress drop necessarily reflect material properties on the fault; variations in past stress history may also contribute to rupture heterogeneity (e.g. Scholz & Aviles 1986). The presence of a characteristic scale in the Shumagin data (dividing regions of low from high \( D_N \)) suggests that heterogeneous fault properties may be required though, as it is difficult to envision purely dynamical processes creating such characteristic scales.

Rupture heterogeneity is often observed for large earthquakes. For example, Kikuchi & Fukao (1987) observed large events \( (M_l > 10^8 \text{Nm}) \) to be composed of many small subevents, which do not completely provide the high-period moment but which have stress drops on an order of magnitude larger than the event as a whole. Similar behaviour is observed in the largest events seen here. Seismograms for the large event 21 show at least three distinct pulses that appear to be distinct subevents, while those for smaller events appear simpler (Fig. 11). Although we were unable to find adequate EGFs for these 1993 events, crude estimates based on total pulse duration at SAN for the subevents (uncorrected for attenuation) suggest \( M_l \) of 2.0, 3.6 and \( 2.1 \times 10^{18} \text{Nm} \) for subevents 1, 2 and 3, corresponding to \( D_N = 0.50, 0.46 \) and 0.40 \( (\Delta \sigma = 25, 30 \) and 46 MPa). Because path corrections were not made these durations are probably an upper bound, but at the dominant frequencies in the SAN displacement record \( (-1 \text{Hz}) \) propagation effects should be small. These subevents show \( D_N \) comparable to that observed for small events, not large events, and show that the differences seen between different events are also reflected in the complexity of large earthquakes.

If taken literally, these arguments imply that only a fraction of the fault surface is well coupled at any time. The simplest asperity models are of patches that rupture at a high stress drop, \( \Delta \sigma_2 \), surrounded by weak material with negligible stress drop. The arguments above imply that small earthquakes have stress drops of \( \Delta \sigma_2 \) on strong patches while larger earthquakes rupture both the strong patches and surrounding materials, giving overall stress drops \( \Delta \sigma_1 \). The Shumagin data imply \( \Delta \sigma_1/\Delta \sigma_2 \sim 50 \) by comparing large to small earthquakes. The heterogeneity of single events is also consistent (Fig. 11). These results suggest that the strong patches make up only a small fraction of the fault area. The actual fraction of the rupture area inferred to be strong depends upon rupture geometry, as rupture complexity can strongly affect the relationship between pulse duration and rupture area (Boatwright 1984). Nevertheless, given the ~50x difference in stress drop and ~4x difference in inferred rupture lengths between large and small events (\( D_N \)) these data may suggest that as little as 1–10 per cent of the interplate thrust may contribute much of the moment release. Because the regions between...
asperities are likely to contribute some to the overall seismic moment these values are probably underestimates.

An alternative is that large values of $D_\alpha$ do not represent larger rupture areas but rather slower rupture velocities. We have arbitrarily assumed a rupture velocity near the Rayleigh wave velocity (0.96), to normalize $D_\alpha$ and to calculate $\Delta\sigma$. One might expect slower rupture velocities on the stable parts of the fault that do not normally slip seismically, either for dynamic reasons or because $\beta$ is anomalously low along weak parts of the fault zone. Changes in rupture velocity will affect $D_\alpha$ linearly, to first order, while $\Delta\sigma$ depends upon the cube of rupture velocity (complex rupture can alter the relationship between $r$ and $D$ in other ways as well; see Boatwright 1984 for a more complete discussion). A 50 per cent reduction in rupture velocity for the largest events would account for most of the variation in $D_\alpha$. More realistically, slow rupture velocities in low-stress areas may act in concert with real stress drop variations to explain the severe differences seen between large and small events.

**Characteristic dimension of faulting**

The change in scaling behaviour we observe between larger and smaller earthquakes is difficult to reconcile with fractal geometry seen on many fault surfaces (e.g. Power et al. 1987). We observe differences between earthquakes larger and smaller than $\sim 5 \times 10^{18}$ Nm, corresponding to a characteristic slip surface size of 3-15 km. The scaling change either corresponds to a physical mechanism of fault behaviour, not previously identified, or to a structural scale length that is inherent on the rocks entering the subduction zone. For example, the frictional yield strength of the fault somehow might act as a natural scale length of the system, in that it sets a material-dependent maximum value on how much fault stress can vary. It is not clear how the yield stress translates into a scale length, but it does provide a limit on the values of shear stress that may affect small events more than large ones.

More likely, the change in scaling properties reflects a scale of material variations in the fault, such as indicated by seismicity variations. Seismicity along the plate interface shows lineations in roughly the direction of plate convergence, at similar scale lengths (Fig. 12). These lineations are reminiscent of inferred seamount tracks seen in bathymetry of accretionary prisms (von Huene et al. 1995) and may also be related to the topography of the downgoing plate. The lineations occur at 5-20 km spacing, similar to the characteristic fault dimension. Hypocentral accuracy is in the order of 1-3 km (Abers 1922a, 1994) so small features might not be recognized. Nevertheless, a characteristic dimension of 5-20 km is suggested.

Large features such as ridges, seamounts and fracture zones can have a profound effect on subduction dynamics (Kelleher & McCann 1977), and it is possible that smaller structures do as well. The observed characteristic dimension of earthquake scaling is close to the $\sim 10$ km characteristic dimensions seen for abyssal hill topography (Gilbert & Malinverno 1988; Goff 1991), and may reflect the influence of inherited sea-floor roughness on the fault zone. Abyssal hill spacing could control the distribution of geometrical irregularities on the lower side of the interplate thrust fault.

![Figure 12. Seismicity along the interplate thrust zone, relocated by Abers (1992a, 1994). Circles show events, and triangles show stations. The figure is aligned in the direction of plate motion on the vertical axis (azimuth of 151°; DeMets et al. 1990). Depths of events are all between 25 and 50 km, with earthquakes more than 10 km from a best-fitting plane eliminated (compare Fig. 2). Small arrows indicate possible linear features that could represent the subduction of rough topography and consequent gouging of the fault zone in the plate motion direction.](https://academic.oup.com/gji/article-abstract/123/1/41/571199)

The horsts and grabens developed during plate flexure seaward of trenches provide a second source of topographic variations on the downgoing plate and also have 10-15 km characteristic dimensions (Hilde 1983). Sediments may play a role, as they may accumulate in the valleys between abyssal hills or in grabens and may be preferentially scraped off the ridges, so that hard-rock contact is limited to the tops of abyssal hills. Basement structures are obscured by sediments in the Alaska peninsula region (Hilde 1983), but short-wavelength variations in depth to the detachment surface can be seen in seismic profiles of the accretionary prism (Lewis et al. 1988). Although speculative, it seems possible that high-stress drop regions correspond to the high parts of subducted topography at characteristic scales of 3-15 km.

**Magnitudes of stress drops from other regions**

Recent static stress-drop studies using empirical Green's function techniques show a broad and variable range of stress drops, many of which are about 30-50 per cent or our values but generally in the 10-30 MPa range (e.g. Frankel et al. 1986; Mori & Frankel 1990; Hough et al. 1991; Xie et al. 1991). Most of those studies are based on small events ($M_0 = 10^{12}$ to $10^{15}$ Nm), and hence would be unable to detect changes in scaling properties at the size ranges observed here. All of these studies examined shallow crustal events, and the difference between the stress drops of those studies and our results may reflect a different behaviour of deeper subduction zone earthquakes. For example, the shear modulus used here is 50-70 per cent larger than values commonly used in upper-crustal studies, and would give
comparable larger stress drops for the same scaling of slip to fault length. Some low stress drops obtained for small earthquakes may be affected by resolution problems analogous to those discussed here, and observations of apparently constant fault radius at small size (i.e. constant duration) should be examined with care.

As with the Shumagin results, many previous studies of small earthquakes occasionally show events having stress drops up to 100 MPa (McGarr 1984; Gariel, Archuleta & Bouchon 1990; Hough et al. 1991). These results may reflect localized stress concentrations, for example at geometrical asperities. Hence, our results do not represent unprecedented high values but merely a high range. By contrast, results from large earthquakes (\(M_0 > 10^{18} - 10^{19}\) Nm) in interplate settings rarely show stress drops greater than 10 MPa (e.g. Kanamori & Anderson 1976). Ekström & Engdahl (1989) find \(D = 2.2 \times 10^{-9} M_0^{1/3}\), for \(M_0\) in dyne centimetres, from 49 events in the Central Aleutians, equivalent to \(D_N = 2.7\). Their relationship is consistent with the largest three earthquakes studied here, but overpredicts durations by a factor of 3–5 for the small events (Fig. 9). Their study is based on earthquakes with \(M_0 \geq 10^{17}\) Nm, or larger than most of the earthquakes that we observe with small \(D_N\), so it is unclear whether an inconsistency exists between their results and ours.

### Stress-drop variation with depth

Laboratory work suggests that stress drops should increase with normal stress on faults (e.g. Scholz et al. 1972). Most work has been done at pressures lower than are relevant for subduction zone earthquakes, and depth dependence of stress drops are not often apparent in seismic data (McGarr 1984). However, the tendency of rupture of large earthquakes to nucleate at the base of the seismogenic zone is evidence for increasing stress drops with depth (Das & Scholz 1983). From our observations, no clear relationship can be seen between the mean \(D_N\) and depth, particularly below 30 km depth (Fig. 13). To some extent the upper bound on stress increases with depth at least in the upper half of the thrust zone (stippled line, Fig. 13(b)). A similar observation was made by Fletcher et al. (1984) for aftershocks of the Oroville, California earthquake, who argued that the static frictional stress sets an upper bound on stress drops. At a given normal stress, stress drops will not exceed the difference between static frictional stress and final stress on the fault after failure, although they may be less; both static and final stresses are likely to increase with normal stress. Our data support such a relationship only at depths shallower than 30 km. As depth increases, ductile processes might become more important, and the pressure dependence of simple frictional behaviour could lessen. Given the lack of obvious depth dependence of stress drop, it is worth considering possible mechanisms for stick-slip behaviour that do not require shear stresses to increase with normal stress. For example, the breakdown of hydrous phases with depth in subduction zones could lead to elevated fluid pressures that could reduce effective normal stresses; other possibilities certainly exist.

Previous interpretations of large increases in stress drop with depth in the Shumagins (House & Boatwright 1980; Taber & Beavan 1986) may reflect the same moment-duration scaling that we observe here rather than depth-dependence of stress drop. The events with very high stress drops (~100 MPa) at the base of the thrust zone, studied by House & Boatwright (1980), fall in our range for the smaller size of events. The shallow 1983 February 14 mainshock studied by Taber & Beavan (1986) is similar in size and duration to event 2 here. Hence, the observation of a low stress drop (in this case dynamic stress drop) may be a consequence of the same complexity argued here for larger earthquakes.
CONCLUSIONS

By using a new empirical Green’s function technique based on damped time-domain inversion, pulse durations were obtained for 20 moderate-sized earthquakes scattered over the main thrust zone in Shumagin region. The results from the deconvolution method compare favourably with those from direct inversion methods. Pulse duration increases with the size of events from 0.1 to 10 s over the range of $M_D$ from $10^{14}$ to $3 \times 10^{16}$Nm. The three largest events ($M_D > 5 \times 10^{16}$Nm) have the three longest normalized durations, and give fault lengths that are similar to or slightly smaller than inferred from the sizes of their early aftershock zones. The normalized durations for smaller events are larger than those for larger events by factors of five to 10. Earthquakes at the base of the thrust zone are not characterized by unusually high stress drops, nor are stress drops in the shallow part of the main thrust zone unusually low. A weak depth dependence on the envelope of stress drops with depth is observed in the upper half of the thrust zone. A moment dependence of $D_N$ is the most striking feature of our measurements, in which the largest events ($M_L > 6$) have several times longer $D_N$ (10–100 $\times$ higher stress drops) than smaller events. These data suggest that numerous small patches are capable of generating high stress drops within a generally weak interplate thrust fault. Small events rupture only strong patches near the frictional yield strength of the fault, while larger events rupture both strongly coupled and weakly coupled zones. Small events rupture patches generally less than 3–5 km in size, while large events have characteristic scales larger than 15–20 km. These dimensions are comparable to characteristic dimensions found in sea-floor topography, and suggest that the subducted sea-floor may be revealed in the scaling behaviour of interplate thrust earthquakes.

ACKNOWLEDGMENTS

We appreciate the efforts of many people who contributed to the establishment and operation of the East Aleutian Seismic Network. D. Johnson provided and helped decipher much of the instrument calibration information needed to make sense of the digital data. Strong motion data for the 1993 events were provided by National Center for Earthquake Engineering Research strong motion program. The approach and ideas developed through conversations with L. Gilbert, S. Hough, K. Jacob and C. Scholz. A. Lerner-Lam pointed us toward time-domain deconvolution, and S. Hough provided the original fitting program used to determine seismic moments. We thank J. Boatwright, W. Menke and A. Lerner-Lam for helpful reviews. The research reported in this paper was supported by USGS awards 14-08-0001-G1981 and 1934-92-G2200, and partially supported by the National Science Foundation grant EAR-9104158. Lamont-Doherty contribution number 5333.

REFERENCES


If \( g_k \) in eq. (1) is a true discrete Green's function then \( s_k \) is the far-field source pulse observed at station \( k \) from the large event. Causality conditions \( g_k(t) = s_k(t) = 0 \) for \( t < 0 \) are enforced by the bounds on the summation in (1); \( 0 \leq j \leq i \). For each pair of seismograms \( \{ h_i, g_j \} \), eq. (1) provides \( n_j \) linear equations for the \( m \) values in \( s_k \), where \( m \leq n_j \) (there are no equations that involve elements of \( s_k \) with \( j > n_j \) because \( g_{k}^{j》 = 0 \). Many rupture models predict that the source pulse \( s_k \) will vary in shape with wave path, as is sometimes observed (e.g. Frankel et al. 1986). However, several source parameters such as duration and seismic moment are not directional properties, and to measure them it is useful to assume that \( s_k = s \) is the same for all \( k \). This assumption increases the number of equations (1) to \( P = n_1 + n_2 + \ldots + n_K \) without increasing the number of parameters to be determined \( (m) \). Thus, (1) can be written as

\[
\begin{bmatrix}
\mathbf{h}_1 \\
\mathbf{h}_2 \\
\vdots \\
\mathbf{h}_K
\end{bmatrix} = \begin{bmatrix}
\mathbf{G}_1 \\
\mathbf{G}_2 \\
\vdots \\
\mathbf{G}_K
\end{bmatrix} \mathbf{s}
\]

which reduces to eq. (2). Each submatrix of \( \mathbf{G} \) takes the

\[
\begin{bmatrix}
\mathbf{h}_1 \\
\mathbf{h}_2 \\
\vdots \\
\mathbf{h}_K
\end{bmatrix} = \begin{bmatrix}
\mathbf{G}_1 \\
\mathbf{G}_2 \\
\vdots \\
\mathbf{G}_K
\end{bmatrix} \mathbf{s}
\]
form

\[ \mathbf{G}_k = \begin{pmatrix} g_k^0 & 0 & \cdots & 0 \\ g_k^1 & g_k^0 & \cdots & 0 \\ \vdots & \vdots & \ddots & \vdots \\ g_k^n & g_k^{n-1} & \cdots & g_k^0 \\ \end{pmatrix} \]  \quad (A2)

Although the inversion of eq. (2), (A2), appears well conditioned, even for a single waveform pair, in practice the noise in observations and the band-limited nature of seismograms make the system of equations ill conditioned at best (e.g. Sipkin & Lerner-Lam 1992). Time-domain stabilization is done in a natural way by directly inverting matrices like (A2) using inverse techniques. For Gaussian uncertainties, the maximum likelihood solution to the linear stabilization is done in a natural way by directly inverting seismograms make the system of equations ill conditioned at

theoretical relationship (2), and contains uncertainties due (e.g. Tarantola & Valette 1982). Here, \( \mathbf{C}_s \) is the \textit{a priori} covariance of \( \mathbf{s} \) and \( \mathbf{C}_{sh} \) is the \textit{a priori} data covariance matrix. The matrix \( \mathbf{C}_T \) is the \textit{a priori} covariances of the theoretical relationship (2), and contains uncertainties due to differences between \( \mathbf{g} \) and the true Green's functions. Formal errors and resolution are calculated in the usual way (Tarantola & Valette 1982).

Unlike other deconvolutions where the transfer function \( \mathbf{g} \) is well known (instrument response removal, etc.), differences between the EGF and the true Green's function can be large and hence \( \mathbf{C}_T \) can be large. At high frequencies, large phase differences may exist between the late parts of seismograms \( \mathbf{g}_k \) and corresponding true Green's function, for example by changes in ray paths of scattered phases. Such phase discrepancies will result in amplitude errors with power comparable to the signal variance of \( \mathbf{g} \) (\( \sigma_g^2 \)). Because \( \mathbf{s} \) is expected to have power comparable to \( \sigma_s^2/\sigma_g^2 \), where \( \sigma_s^2 \) is the power in \( \mathbf{h} \), the errors in \( \mathbf{g} \) will result in theoretical uncertainties in \( \mathbf{G}_s \) with variance of order \( \sigma_g^2 \). The \textit{a priori} covariances \( \mathbf{C}_s \) should then scale with \( \sigma_s^2 \). By contrast, the expected data variance \( \mathbf{C}_{sh} \) is generally small compared to the power in \( \mathbf{h} \), because background noise levels must be sufficiently small to allow recording of the much smaller event \( \mathbf{g} \). Thus, \( \mathbf{C}_T + \mathbf{C}_{sh} \approx \mathbf{C}_T \) and scales with the signal power in \( \mathbf{h} \). For simplicity we assume that the data uncertainties are independent, so \( \mathbf{C}_T \) is diagonal and

\[ \{\mathbf{C}_T\}_i = a^2 \sigma_i^2 \]  \quad (A4)

where \( \sigma_i^2 \) is the RMS amplitude of the seismogram \( \mathbf{h}_k \) corresponding to datum \( i \) and \( a \) is some specified constant.

A simple strategy for estimating \( \mathbf{C}_s \), adopted here, is to assume that \( \mathbf{C}_s \) is diagonal with all elements equal to some model variance \( \sigma_s^2 \). We expect that the 'size' of \( \mathbf{s} \) is of the order \( \beta \) where \( \beta \) is the ratio of seismic moments between the two earthquakes, since to first approximation the integral of \( \mathbf{s} \) should give the moment ratio. This gives \( \{\mathbf{C}_s\}_i = (\beta/c)^2 \) for some constant \( c \). The resulting estimator has the form of eq. (3), where for multiple traces from the same event pair

\[ \mathbf{W} = [\mathbf{W}_1 \mathbf{W}_2 \cdots \mathbf{W}_n]^T; \quad \mathbf{W}_k = \sigma_k^{-2} \mathbf{I}_{nk} \]  \quad (A5)

The assigned constant \( \epsilon = a/c \) (eq. 3) scales the regularization.

**APPENDIX B: DIRECTION INVERSIONS FOR PULSE DURATION**

We examine source time functions parametrized by an amplitude \( A \) and duration \( D \):

\[ \mathbf{s} = A \mathbf{m}(D). \]  \quad (B1)

In this case, the \( P \) equations (A2) can be rewritten as eq. (4) which depends upon a model vector \( \mathbf{f}(D) = \mathbf{G}_m(D) \). For Gaussian prior uncertainties, the complete \textit{a posteriori} probability density function (pdf) \( \rho_m \) for the model parameters \{\( A, D \)\} is (Tarantola & Valette 1982)

\[ \rho_m(A, D) = c \sigma_m(A, D) \exp \left\{ -\mathbf{h}^T [\mathbf{h} - A \mathbf{f}(D)]^T \right\} \]  \quad (B2)

where \( \sigma_m(A, D) \) is the \textit{a priori} pdf for \( A \) and \( D \), and \( c \) is the constant that normalizes \( \rho_m \) (the state of null information discussed by Tarantola & Valette (1982) should be constant for this problem). The problem is simplified further by assuming that we have no prior information about \( A \) and \( D \), and set

\[ \sigma_m(A, D) = \text{constant}; \quad A \geq 0 \text{ and } D \geq 0; \]

\[ = 0 \text{ otherwise.} \]  \quad (B3)

In general, we are more interested in \( D \) than in \( A \), so we calculate the marginal pdf of \( D \). Integration of (B2) over \( A \), using (B3), gives

\[ \rho_m(D) = \int_{-\infty}^{\infty} \rho_m(A, D) dA = c_3 \exp \left\{ -R - S^2/T \right\} \]  \quad (B4)

where \( c_3 \) is the scaling constant and

\[ R = \mathbf{h}^T (\mathbf{C}_{hh} + \mathbf{C}_T)^{-1} \mathbf{h}, \]  \quad (B5)

\[ S = \mathbf{h}^T (\mathbf{C}_{hh} + \mathbf{C}_T)^{-1} \mathbf{f}(D), \]  \quad (B6)

\[ T = \mathbf{f}(D)^T (\mathbf{C}_{hh} + \mathbf{C}_T)^{-1} \mathbf{f}(D). \]  \quad (B7)

Eq. (B4) provides a direct analytical form for \( \rho_m(D) \) which is then calculated for a range of possible values of \( D \); the constant \( c_3 \) is determined numerically to give \( \int \rho_m(D) dD = 1 \). This result holds for any simple source parametrization of the form (B1). The extension to a complex source pulse model is fairly straightforward, by treating \( D \) in the above expressions as an array of rupture parameters rather than a single scalar. In practice, one additional parameter is necessary, the arrival time difference \( \delta \) between the two events, which gives the expected start time of the source pulse and trades off with \( D \).

The primary assumption is that of Gaussian statistics, although the \textit{a priori} covariance matrices \( \mathbf{C}_T \) and \( \mathbf{C}_s \) are not required to be diagonal so that waveform autocorrelations can be accounted for explicitly. However, if they are diagonal, it can be shown from (B4)–(B7) that \( \rho_m(D) \) scales to the maximum cross-correlation between \( \mathbf{h} \) and \( \mathbf{f}(D) \). In all inversions shown here, we assume that \( \mathbf{C}_{dd} + \mathbf{C}_T \) is diagonal with each entry equal to the sum-squared amplitude of the seismogram corresponding to that row.

© 1995 RAS, GJI 123, 41–58