The nucleation and rupture process of the 1981 Gulf of Corinth earthquakes from deconvolved broad-band data

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SUMMARY

Source parameters of the largest three normal faulting earthquakes (M6.6, 6.3, 6.4), in the 1981 Gulf of Corinth (Greece) sequence are determined using deconvolved broad-band data (recorded by arrays and single stations) and a 2-D finite source model. Such a model enables the spatial extent, rupture velocity and stress drop of the earthquakes to be determined and geological observations of surface slip can be included as a further constraint on the waveform modelling. All three earthquakes were shallow (<10 km) with low stress drops (<30 bars), and exhibited source complexity. The correspondence between the complexity of the earthquake sources and that of the mapped fault breaks implies that the segmentation of surface faulting in Greece is representative of faulting at depth. Tiny initial pulses which correlate across the arrays are seen in the seismograms from the Gulf of Corinth earthquakes at most stations. These initial subevents (<1 per cent of the total moment) are interpreted as the breaking of small asperities which initiated the main rupture and are used to constrain the attenuation correction (t* = 0.2 s).

Key words: Aegean, body waves, 2-D model, earthquake, rupture and nucleation.

1 INTRODUCTION

The Aegean region, including Greece and the Aegean Sea, is one of the most actively deforming sedimentary basins in the world (e.g. Jackson & McKenzie 1988). Rapid north–south extension at about 7 cm yr⁻¹ (McKenzie 1978) is occurring behind the Hellenic subduction zone, producing high seismicity on a complex distribution of normal faults. Previous studies of Aegean seismicity have determined basic earthquake source parameters such as fault-plane solutions, depths and seismic moment using point-source approximations to model long-period (LP, 20 s) seismic body waves (e.g. Jackson et al. 1982; Kim, Kulhánek & Meyer 1984; Taymaz, Jackson & McKenzie 1991). These results are useful in determining the regional stress field and seismic strain rate, but are less informative about the rupture processes of the earthquakes themselves. For example, it has been suggested that Aegean earthquakes have low stress drops (about 10 bars, Kim et al. 1984) but these values are poorly constrained. It is now well known that a much greater resolution of the earthquake rupture process can be obtained from body waves containing a broad spectral bandwidth than from conventional long- and short-period (SP) seismograms (e.g. Choy & Boatwright 1981, 1982).

Broad-band seismograms containing wavelengths of the same size as earthquake faults show details of the earthquake sources not recorded by the LP waves (e.g. Choy & Kind 1987). Also, Der, Shumway & Hirano (1991) demonstrate that most LP waveform modelling studies have considerably less resolution than is typically claimed. A good model of the waveform filtered around 20 s (LP) may be a poor fit at the periods at which the source radiates most of its energy. Modelling of body waves containing the frequencies 0.1–5 Hz is more accurate, as the maximum energy release by moderate-sized earthquakes is in this range.

In this study teleseismic broad-band body waves are used to investigate the source mechanisms of three recent, normal faulting earthquakes (M6.6, 6.3 and 6.4, respectively) which occurred in the Gulf of Corinth, 1981 (Fig. 1 and Table 1). These earthquakes are representative of the seismicity in the Aegean back arc basin. They occurred on major basin-bounding faults which control the topography. The rupture processes of these earthquakes are determined here by fitting the P waves with a 2-D source model enabling us to investigate source directivity and include any mapped surface faulting as an additional constraint on the waveform modelling. An important result of this method is that the relationship between the often complex fault breaks mapped at the surface and the fault rupture at depth can be
determined. The acceptance of the idea of characteristic earthquakes rupturing the same piece of fault repeatedly, their size controlled by fault geometry, has led to seismic hazard estimates based on the pattern of faults observed at the surface. It is clearly important to determine whether fault segmentation observed at the surface is representative of complex fault rupture at depth, or whether it is just caused by near-surface heterogeneity.

2 DATA AND PROCESSING

The principal data used in this study are from two seismometer arrays: Gauribidanur (GBA), India, and Yellowknife (YKA), Canada. Both consist of 20 SP vertical instruments. When the location of the earthquake is known, the recordings by the array seismometers (channels) can be summed, allowing for the appropriate time delays at each instrument. This optimal summation significantly increases the signal-to-noise ratio over that at an individual instrument (Fig. 2).

The SP instrument response is removed from the summed records to produce displacement seismograms with a flat frequency response below 10 Hz (following Douglas, Richardson & Hutchins 1990; Douglas, Sheehan & Stewart 1992). Wiener filtering is used to remove low-frequency noise, with an average passband of 100 per cent at 0.15 to 5-6 Hz and 50 per cent by 7-8 Hz. Recovery of longer period energy from the array recordings is aided by their response falling off much less steeply than that of Global Digital Seismograph Network (GDSN) or World Wide Standard Seismograph Network (WWSSN) instruments (Douglas et al. 1992). Also, the response fall-off below 1 Hz is compensated for, to a significant extent, by the rapid rise in amplitude of the source spectrum until the corner frequency.

Table 1. Hypocentral parameters of the three Gulf of Corinth earthquakes included in this study as reported by the ISC. The depths (D) are the results of this study, and the latitudes and longitudes are those determined by Jackson et al. (1982).

<table>
<thead>
<tr>
<th>No.</th>
<th>Date</th>
<th>Time</th>
<th>Lat.</th>
<th>Long.</th>
<th>D</th>
<th>M₃</th>
<th>m_b</th>
</tr>
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<tbody>
<tr>
<td></td>
<td></td>
<td>(Hr Min S)</td>
<td>(°N)</td>
<td>(°E)</td>
<td>(km)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>24 Feb 81</td>
<td>20 53 37</td>
<td>38.10</td>
<td>22.84</td>
<td>5</td>
<td>6.6</td>
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</tr>
<tr>
<td>2</td>
<td>25 Feb 81</td>
<td>2 35 53</td>
<td>38.14</td>
<td>23.05</td>
<td>5</td>
<td>6.3</td>
<td>5.7</td>
</tr>
<tr>
<td>3</td>
<td>4 Mar 81</td>
<td>21 58 7</td>
<td>38.18</td>
<td>23.17</td>
<td>3</td>
<td>6.4</td>
<td>5.8</td>
</tr>
</tbody>
</table>
frequency is reached. Douglas, Stewart & Richardson (1984) and Douglas et al. (1990, 1992) show periods of several 10s of seconds recovered from the array SP recordings. The excellent agreement between the displacement seismograms we obtain from the YKA SP channels and the broad-band seismograms recorded at the same site (integrated velocity channel) confirm the success of our deconvolution method (Fig. 3).

Stewart & Douglas (1983) demonstrate that phaseless seismograms from which the phase shifts of the recording system have been removed show more clearly the form of the seismic pulses than do conventional broad-band seismograms. Such seismograms are preferred here since we are interested in the shape of the waveform rather than its arrival time. Unfortunately the phase shift removal produces artificial, spurious precursors to some of the waveforms (e.g. Douglas et al. 1992). These can, however, easily be recognized and should be ignored. All three earthquakes considered here are normal faulting events and all tectonic first motions are negative. The processing generated precursors are always of opposite sign and so are all positive for the events considered here.

Having successfully obtained relatively broad-band displacement seismograms for the three earthquakes from the array data, SP recordings from the GDSN were converted to displacement in the same way. Despite the steeper fall-off of the GDSN instruments below 1 Hz Bezzeghoud, Deschamps & Madariaga (1986) were able to extract signal out to 0.066Hz for the Gulf of Corinth earthquakes. The displacement records we deconvolved from the GDSN SP do not appear to be lacking in low-frequency energy compared with the array seismograms. They are therefore included in the modelling to increase azimuthal coverage.

To correct the seismograms for intrinsic attenuation we use the attenuation operator \( b(0) \) (Carpenter 1966a), with phase shifts after Futterman (1962), where \( b(0) \) can be expressed in the form \( (1+a^2t^2)/2 \) and \( a^2 \) is the product of the traveltime and the average value of the reciprocal of the specific quality factor \( Q \) on the path from the source to the station. We observe small, high-frequency pulses (e.g. Fig. 4, arrival I) preceding the main energy release of the three Gulf of Corinth earthquakes which we interpret as precursory subevents initiating the main rupture (see below). Intrinsic attenuation acts to broaden seismic pulses by preferentially removing high-frequency energy, and therefore the rise and fall times of these initial pulses (0.2-0.6s, Abercrombie 1991) can be used to estimate a maximum value of \( a^2 \), i.e. that needed to broaden a delta function to the recorded pulse width (Stewart 1984). A value of \( a^2 = 0.2 \) s (corresponding to \( 1/Q = 0.0003 \), averaged over the ray path) is chosen as being well within the upper bound and is applied to all the displacement seismograms. Stewart (1984) also finds that some transmission paths have an average attenuation factor considerably lower than the \( a^2 \) of 1.0s which is often used when modelling seismograms at teleseismic distances. Douglas (1992) determines \( a^2 \) to be 0.15-0.35 between 1 and 10 Hz for nuclear tests recorded teleseismically. Some frequency dependence of \( a^2 \) has been observed (e.g. Cormier 1982) with values similar to those used here above 1 Hz increasing to about 1 s below 1 Hz. The use of a frequency-dependent \( a^2 \) such as that suggested by Choy & Cormier (1986) or Der, Lees & Cormier (1986) would cause a minor trade-off between source duration and attenuation operator. This should not affect our long-period estimate of moment, but might cause our estimate of fault length to decrease by about 10 per cent.

### 3 SOURCE PARAMETERS

#### 3.1 The 2-D finite-dislocation model

The model used in this study is a kinematic 2-D finite-dislocation model of the type suggested by Savage (1966). Previous LP waveform modelling of earthquakes in the Aegean has assumed a point-source (e.g. Soufleris & Stewart 1981; Jackson et al. 1982; Lyon-Caen et al. 1988; Taymaz et al. 1991). This is a valid approximation for moderate-sized earthquakes recorded at such long wavelengths (20s corresponds to about 120 km, and the source size is of the order of 10km). However, this approximation is less good for broad-band data, which include wavelengths (1-100km) of the order of the spatial dimensions of the earthquake source, and so a 2-D model should be used. This type of model also has the advantage of enabling determination of such source parameters as fault size and amount of slip.

In the Savage model, the earthquake sources is assumed to be a double couple of finite size, and fracture is assumed to start at the centre or focus of an ellipse (an adaptation of the Savage model by Douglas, Hudson & Blamey 1972) propagating out at a uniform rupture velocity until it reaches the boundary. Slip occurs instantaneously at each point as the rupture front passes. The final slip distribution is of the form

\[
\text{Slip at a point } = \left(0.1 - \frac{x^2}{r^2}\right)^{1/2},
\]

where \( x \) is the distance of the point from the centre of the fault, and \( r \) that from the edge of the fault to the centre. Such a slip distribution has been reproduced experimentally (Archuleta & Brune 1975), and has also been used.
Figure 4. 1981 February 24 earthquake in the Gulf of Corinth recorded at GBA: (a) the SP sum, (b) the deconvolved displacement, and (c) the amplified displacement P-wave seismograms. Two initial subevents (I and II) can be seen to the main event (M). The first, and smallest, arrival is seen in the magnified displacement record corresponding to the SP onset. As the pulse is about 1 s long, its amplitude is greatly magnified with respect to the longer duration mainshock in the SP recording. It correlates both spatially and temporally across the array so it must have come from the source region. (d) The calculation of the rise and fall times and pulse area. Note that the pulse is negative, with the same polarity as the main event, against a positively trending background of the low-frequency processing generated precursor.

3.2 P-wave modelling

Seismic waveform modelling of such shallow earthquakes as those in Greece is complicated by the interference between the surface reflections ($pP$ and $sP$) and the direct $P$ wave. In this study we model the $P$ waveforms to obtain basic source parameters and to investigate how any source complexity is related to surface rupture patterns. The presence or absence of surface faulting is used to constrain the waveform models, as the surface faulting which they predict must be consistent with that observed. We find, in common with previous studies (e.g. Soufleris & Stewart 1981) that use plane layered observations), and amount of surface slip predicted by the model (Abercrombie 1991). This is not possible with the point-source models previously used to study Aegean earthquakes and is important because it enables surface slip observations to be used as direct constraints on the seismic waveform modelling.
velocity structures, that alterations to these structures (for example, adding the shallow water layer in the Gulf of Corinth) have a minimal effect on the synthetic seismograms. We, therefore, use a simple velocity structure based on seismic refraction work in the Aegean area (Macris & Vees 1977) to model the three earthquakes (Table 2). We make no attempt to model arrivals following the direct and first reflected waves because such arrivals are at least as likely to result from structural heterogeneity as from source complexity. For example, the faulting in the Aegean area is known to be of a tilting block nature (Jackson & McKenzie 1983) and Wiens (1989) and Lundgren, Okal & Wiens (1989) have shown that earth structures consisting of dipping layers can lead to complex reflections and reverberations.

The modelling was carried out by varying the depth and radius and stress drop until a good fit to the observed seismograms was obtained. The rupture velocity was assumed to be 2.1 km s\(^{-1}\) (0.6\(\beta\)), consistent with previous work suggesting low rupture velocities in the region (e.g. Kim et al. 1984), and was only varied if no good fit could be obtained. As in all such forward modelling studies, there is some degree of trade-off between the various model parameters. This is considerably lessened by the extra constraint of predicting surface faulting compatible with that observed. Fig. 5 shows the effect of varying the depth and radius for a single event. The goodness of fit was determined by eye, taking into account the precise shape of the waveform. We tested some of our models with a simple sensitivity analysis based on the peak amplitude and duration of the \(P\) waveform. This analysis (Abercrombie 1991) confirmed that the errors are approximately ±2 km depth and ±2 km radius. Since for any given radius and moment the stress drop is fixed (Eshelby 1957), the stress drop is a parameter which is independent of waveform duration, and so can be used to scale the model of given depth and dimension to fit the peak-to-peak amplitudes at each station. An estimate of the errors in stress drop, and hence moment, can thus be found from the station to station variation. The mean value for all stations, and its standard deviation (±5–10 bars) are quoted in the following for both stress drop and moment. Clearly there are many more sources of error than just these in the particular model, for example, structural heterogeneity path effects and appropriateness of the model, so these errors should probably be considered minimum estimates.

4 GULF OF CORINTH EARTHQUAKE SEQUENCE 1981

There were no foreshocks to the first and largest earthquake of the 1981 Gulf of Corinth sequence on February 24 (6.6 \(M_\text{S}\)) which was followed within a few hours by the second earthquake on February 25 (6.3 \(M_\text{S}\)). The third earthquake in the sequence occurred to the north-east, near Kaparelli, on March 4 (6.4 \(M_\text{S}\)). The earthquake sequence and the surface faulting (up to 1 m of slip) produced are well described by Jackson et al. (1982). Fig. 6 shows the epicentres and surface breaks, most of which involved reactivation of old bedrock scarps responsible for the topography. The aftershocks to the three largest events appear to be concentrated in the hanging wall of the north-dipping faults, and are in the depth range 4–12 km with depth errors ±4 km (King et al. 1985).

Although various authors have worked on these earthquakes, there is still debate over the details of the rupture processes. For example, Kim et al. (1984) and Bezzeghoud et al. (1986) noted possible small precursors to the main \(P\) waveforms in the seismograms, but had too poor data to study them. Also, the relationship between the mapped surface breaks and the faulting at depth is still unclear. Jackson et al. (1982) propose that the February 24 earthquake ruptured an offshore fault and that the February 25 earthquake was responsible for all the faulting observed on the Perahora peninsula, whereas Papazachos et al. (1984) and Vita-Finzi & King (1985) suggest that the February 24 earthquake may have caused the surface break on the peninsula to the west of Psia. More recently, Taymaz et al. (1991) propose that the February 24 earthquake was responsible for all the primary surface faulting on the peninsula, and that the February 25 earthquake ruptured a smaller fault to the east of Skinos. The March 4 earthquake produced two clear, offset surface breaks, but no previous study has been able to determine whether rupture at depth occurred on two faults or whether a simple, single rupture took place at depth in the crust, and near-surface structural heterogeneities are responsible for the more complicated surface faulting.

The relocations of these earthquakes by Jackson et al. (1982) and the focal mechanisms of Kim et al. (1984) are used in the modelling described below.

4.1 1981 February 24 earthquake (6.6 \(M_\text{S}\))

Two distinct arrivals (II and M) are observed in all seismograms of this earthquake (Fig. 7). A very small pulse (I), preceding these and corresponding to the onset on the SP record, is observed at the GBA array (Fig. 4). Although it is only just observable above the noise level, it has clearly travelled from the source region because it appears in the summed trace, and is also seen at the quietest GDSN stations (see Table 3). All three observed arrivals have negative polarity at all stations, and so probably have similar focal mechanisms. The relative arrival times of the three pulses are accurately picked (± approximately 0.05 s) using both the SP recordings and the displacement seismograms at all the stations where they are observed. The first pulse (I) is too small to model, but the area of the pulse on the broad-band seismograms can be used to estimate its moment using eq. (2), (Pilant 1979) as it is of too short a duration for the surface reflections to interfere with the direct \(P\) pulse.

\[
M_\text{w} = \frac{4\pi \rho a^2}{2G(\Delta) U_{\text{ew}}} \quad \text{area.} \tag{2}
\]

The density \(\rho\) and \(P\)-wave velocity \(a\) are taken from the crustal structure used, the mean radiation pattern for \(P\) waves \((U_{\text{ew}})\) is 0.52 (Aki & Richards 1980) and \(G(\Delta)\) is a factor to account for geometrical spreading (Carpenter

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**Table 2.** Velocity structure used for body-wave modelling.  

<table>
<thead>
<tr>
<th>(P) Velocity</th>
<th>(S) Velocity</th>
<th>Density</th>
<th>Thickness</th>
</tr>
</thead>
<tbody>
<tr>
<td>(km/s)</td>
<td>(km/s)</td>
<td>(kg/m(^3))</td>
<td>(km)</td>
</tr>
<tr>
<td>6.0</td>
<td>3.5</td>
<td>2.8</td>
<td>35.0</td>
</tr>
<tr>
<td>8.1</td>
<td>4.7</td>
<td>3.3</td>
<td>-</td>
</tr>
</tbody>
</table>

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Figure 5. 1981 February 24: circular source models of the largest subevent (solid lines) in the displacement P-wave seismogram (dashed line) recorded at GBA (see Table 4). (a) Varying depth, radius fixed at 12 km, (b) varying radius, depth fixed at 5 km. The inflection on the negative P pulse is interpreted as a small event (I in Fig. 4) prior to the main rupture.

1966b). The results for this and for the first subevents observed in the February 25 and March 4 earthquake seismograms are shown in Table 3(a). The seismic moment of this first subevent is found to be \(4.7 \times 10^{15}\) Nm (within about a factor of 10).

The second (II) and third (M) arrivals are modelled using circular sources with the same epicentre and focal mechanism. The parameters are varied until a good fit is obtained (Fig. 7) and the source parameters of the preferred model are listed in Table 4. The combined seismic moment of the three subevents is \(6.0 \pm 2.4 \times 10^{16}\) Nm. The earthquake grew at an increasing rate: 0.1 per cent of total moment was released in the first subevent, 5.3 per cent in the second and 94.6 per cent in the mainshock. The subevents I and II are interpreted as the breaking of small asperities which dynamically triggered the main rupture.

The first two subevents are located relative to the third (mainshock), using the time delays between the three arrivals at the different stations. The depth is fixed at various values, and 5 km produced the lowest residuals for both events. The location results are shown in Table 3, together with the errors. The location of subevent II is better constrained than that of I because more stations (seven, including BCAO to the south) are used. Subevent II is, within errors, colocated with the onset of the main energy release and subevent I appears to have initiated between 2 and 10 km to the west.

4.2 1981 February 25 earthquake (6.3 \(M_g\))

The waveforms of this earthquake are relatively simple and are well fitted by a single, circular source (Table 4, Fig. 8) with moment \(1.5 \pm 0.6 \times 10^{16}\) Nm. As in the February 24 earthquake, the main energy release was preceded (by about 1 s) by an arrival clearly observed at GBA and two GDSN stations. The seismic moment of this subevent is calculated to be \(2.5 \times 10^{15}\) Nm (0.2 per cent of the total moment), but there are too few recordings to attempt a location.

4.3 1981 March 4 earthquake (6.4 \(M_s\))

The seismograms recorded from this earthquake are significantly more complex than those from the previous two earthquakes in the sequence. The direct negative P pulse
Source modelling of Aegean earthquakes

Figure 6. Map of the eastern end of the Gulf of Corinth, showing the locations (Jackson et al. 1982) and focal mechanisms (Kim et al. 1984) of the three earthquakes in the 1981 sequence. Also marked are the mapped bedrock surface breaks on the Perahora peninsula and to the south of Kaparelli. The offshore fault on which the February 24 earthquake may have occurred is shown in the Gulf; it either changed strike at its eastern end, or continued onshore. After Jackson et al. (1982).

shows an inflection at most stations, and the surface reflections are much higher frequency than predicted by a circular model. These higher frequency surface reflections are also noted by Bezzeghoud et al. (1986). A simple circular or elliptical source model is a poor fit to the broad-band seismograms from this earthquake even though it is possible to fit the lower resolution LP data with a single point-source model (e.g. Jackson et al. 1982; Kim et al. 1984).

We attempted to model this earthquake by combining two elliptical sources in different orientations at varying time delays, (their size and depth constrained by the mapped surface faulting) to determine if interference could produce the complex surface reflections observed. The best fit is obtained using two ellipses with major axes perpendicular to strike, rupturing from the centre (Fig. 9) This fit is not considered excellent, but is a significant improvement on a simple circular source.

In the model the Doppler effect is partly responsible for the radiated pulse shapes. When the rupture propagates downwards, as in the preferred model, the direct P pulse is of increased frequency and amplitude, and the surface reflections, propagating in the opposite direction to the source, experience the reverse effect.

Without detailed knowledge of local structural heterogeneity a reliable, unique fit to the complex waveforms radiated by this earthquake cannot be achieved, but all indications, including the high (42 per cent) non-double couple component in the CMT solution (Dziewonski & Woodhouse 1983), point to it being a complex earthquake, consisting of two similar sized sources, and that the pattern of faulting mapped at the surface extends to depth.

Again, the main energy release is preceded (by about 1 s) by an arrival clearly observed at both arrays and three GDSN stations (Table 3).

5 DISCUSSION

In Table 5 the source parameters presented here are compared with those determined by previous studies of the Gulf of Corinth earthquakes. The depth range of nucleation of the three earthquakes determined in this study is 3–5 km,
and the mean depth of maximum moment release (from the model fault slip distribution) is approximately 5 km. The maximum depth of seismic faulting is 6-15 km which is in good agreement with the maximum depth of hypocentres in the back arc area, and confirms the seismogenic zone in the Aegean to be approximately 10 km thick. The depths of maximum moment release determined in this study are close to those found by Bezzeghoud et al. (1986) who also used broad-band data (Table 5). They are systematically shallower than the depths determined by LP studies (e.g. Jackson et al. 1982; Kim et al. 1984), probably because of the lower resolution of LP data and because the exact nature of an LP depth is unknown.

Although most earthquakes are thought to nucleate near the base of the brittle crust and rupture upwards (Das & Scholz 1983), the March 4 earthquake exhibits directivity effects consistent with downward rupture propagation. Aftershock activity is higher at the eastern end of the Gulf of Corinth sequence, and deepest events are at that end (King et al. 1985). This aftershock distribution provides tentative support for downward propagation in the March 4 1981 earthquake. Earthquakes which propagate downwards have been identified in Chile and the CIS (Douglas et al. 1990, 1992) and in California (Dreger & Helmberger 1991) and so fractures which propagate preferentially to increasing depths may not be uncommon.

The stress drops of all three earthquakes are relatively low, in the range 5-30 bar, confirming the suggestion of Kim et al. (1984). A possible explanation for this is related to the residual stresses and friction on the faults. Sibson (1974) suggests that lower stresses are needed to initiate sliding on normal faults than on any other type. In the Aegean area, the high strain rate implies a large (horizontal) extensional force producing a low confining pressure on the normal faults, and hence, low friction. If less stress build up is required to cause an earthquake, then lower stress drops during rupture might be expected. Cocco & Rovelli (1989) report evidence for significantly lower stress drops during normal faulting earthquakes than during thrusts, in Italy.

We assumed a relatively low rupture velocity (Vr = 0.6b) for the modelling, and were able to fit two of the earthquakes without need to vary it. It would be possible to fit the waveforms using a higher rupture velocity, but increasing Vr to 0.7b would require the model stress drop to half and the dimensions to be increased by 2-5 km. The surface faulting predicted by such models would be incompatible with that observed, and such low stress drops are unlikely. The February 24 earthquake was fitted better with the slightly lower rupture velocity of 2.0 km s⁻¹, increasing our confidence in the hypothesis that Aegean earthquakes have relatively low rupture velocities.

The seismic moments determined in this study for the three earthquakes in the Gulf of Corinth are comparable with those obtained from previous LP studies (Table 5), and a factor of 2-3 lower than the published CMT solutions (Dziewonski & Woodhouse 1983). These results confirm that our deconvolved displacement data do not lack significant long-period energy. The somewhat higher moment estimates obtained by Bezzeghoud et al. (1986), using a subset of the data used here, probably result from their modelling the entire waveform as source radiation and not taking into account reflections from structural heterogeneity.

Ekström & Dziewonski (1988) find that the moments of earthquakes in the Aegean area are low with respect to their surface-wave magnitudes. The failure of CMT solutions to
model source complexity was suggested by Main & Burton (1989, 1990) as a possible cause of low moments for Aegean earthquakes. The above results and discussion show that this is clearly not the case. The cause of the anomaly can best be explained by \( M_s \) being overestimated for earthquakes in this region (Abercrombie 1994).

The three earthquakes in the Gulf of Corinth sequence are observed to have initiated with small subevents which were responsible for less than 1 per cent of the total moment release. Examination of the literature shows that a number of moderate and large earthquakes were similarly initiated by small subevents, for example, Chile 1985, M7.8 (Choy & Dewey 1988), Superstition Hills 1987, M6.2 (Wald, Helmberger & Hartzell 1990), Armenia 1989, M6.9, (Haessler et al. 1992), Loma Prieta 1989, M6.9 (Wald, Helmberger & Heaton 1991), and, most recently, Landers 1992, M7.3 (Abercrombie & Mori 1994). Abercrombie & Mori use local observations of the Landers earthquake to show that the two early subevents (M4.4 and M5.6) are indistinguishable from typical earthquakes of such magnitudes. The above observations combine to suggest that dynamic stresses from commonly nucleating small earthquakes may be as important as long-term static stresses in nucleating large earthquakes.

The locations of the subevents and predicted surface breaks for the Gulf of Corinth sequence determined in this study are shown in Fig. 10. We were able to obtain good fits to the waveforms with models which predict surface rupture comparable with that observed. The relative locations and lengths of the surface faults predicted for the February 24 and 25 earthquakes show clearly that the February 25 earthquake cannot have been responsible for all the surface faulting observed on the Perahora peninsula. The February 24 earthquake must have continued onshore and caused the fault breaks to the west of Pisia, as proposed by Papazachos et al. (1984) and Vita-Finzi & King (1985). The change in strike of the surface faulting from west to east of Pisia is also marked by the change in strike (and slip vector) between the February 24 and 25 earthquake focal mechanisms. The model of faulting proposed by Taymaz et al. (1991) cannot be ruled out by the results of this study, but it is not preferred because it fails to account for the surface faulting.

Table 3. (a) Seismic moments and magnitudes of the small initial subevents of the Gulf of Corinth earthquakes, 1981. \( M_s \) is calculated by the standard method (Gutenberg & Richter 1956), and \( M_w \) is the moment-magnitude (Kanamori 1977). The errors quoted are the 95% confidence limits.

<table>
<thead>
<tr>
<th>Earthquake</th>
<th>( M_s ) (x ( 10^{15} ) Nm)</th>
<th>( M_w )</th>
<th>( m_b )</th>
</tr>
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<tr>
<td>24 Feb 1981</td>
<td>4.65 (+46.01 / -4.23)</td>
<td>4.4</td>
<td>5.0 ± 0.5</td>
</tr>
<tr>
<td>25 Feb 1981</td>
<td>2.53 (+52.46 / -2.41)</td>
<td>4.2</td>
<td>4.9 ± 0.4</td>
</tr>
<tr>
<td>4 Mar 1981</td>
<td>9.26 (+16.14 / -5.88)</td>
<td>4.6</td>
<td>4.7 ± 0.4</td>
</tr>
</tbody>
</table>

Table 3. (b) Location results of the two precursory subevents to the 1981 February 24 earthquake. The origin time (T) is relative to the mainshock. Azi is the angle from the subevent epicentre to the mainshock epicentre, Dist is the distance between them, and Res is the sum of the squared residuals (s). N is the number of stations used in the relative location.

<table>
<thead>
<tr>
<th>Event</th>
<th>Depth (km)</th>
<th>Lat. (°N)</th>
<th>Long. (°E)</th>
<th>T (s)</th>
<th>Dist (km)</th>
<th>Azi (°)</th>
<th>Res (s)</th>
<th>N</th>
</tr>
</thead>
<tbody>
<tr>
<td>I</td>
<td>5.0</td>
<td>38.09</td>
<td>22.77</td>
<td>-3.3</td>
<td>6.1</td>
<td>75.9</td>
<td>0.0500</td>
<td>6</td>
</tr>
<tr>
<td></td>
<td>(±4.4 km)</td>
<td>(±3.9 km)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>II</td>
<td>5.0</td>
<td>38.10</td>
<td>22.83</td>
<td>-1.5</td>
<td>0.9</td>
<td>115.5</td>
<td>0.0055</td>
<td>7</td>
</tr>
<tr>
<td></td>
<td>(±1.1 km)</td>
<td>(±1.0 km)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Table 4. Source parameters of the three Gulf of Corinth earthquakes determined in this study. D is the depth of nucleation, \( r \) is the radius, or major \( \times \) minor axes of ellipse, \( D_m \) is the maximum depth reached, \( B \) is the length of the surface break, and \( \ell_m \) is the slip at the centre of the fault. All errors are the 95% per cent confidence limits, and (i) and (ii) denote subevents.

<table>
<thead>
<tr>
<th>Earthquake</th>
<th>( D ) (km)</th>
<th>( r ) (km)</th>
<th>( \Delta \sigma ) (bars)</th>
<th>( V_R ) (km/s)</th>
<th>( M_s ) (total) (x10(^{16} ) Nm)</th>
<th>( D_m ) (km)</th>
<th>( B ) (km)</th>
<th>( \ell_m ) (cm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1. Corinth (i)</td>
<td>5.0</td>
<td>4.0</td>
<td>21.9 ± 10.4</td>
<td>2.0</td>
<td>604 ± 240</td>
<td>13.0</td>
<td>18</td>
<td>70</td>
</tr>
<tr>
<td>24.2.81    (ii)</td>
<td>5.0</td>
<td>12.0</td>
<td>15.8 ± 7.5</td>
<td>2.0</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>2. Corinth</td>
<td>5.0</td>
<td>8.0</td>
<td>13.2 ± 5.4</td>
<td>2.1</td>
<td>154 ± 63</td>
<td>6.0</td>
<td>6</td>
<td>40</td>
</tr>
<tr>
<td>25.2.81</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>3. Corinth (i)</td>
<td>3.0</td>
<td>16.0 × 3.5</td>
<td>15.8 ± 10.2</td>
<td>2.1</td>
<td>134 ± 61</td>
<td>14.7</td>
<td>12</td>
<td>11</td>
</tr>
<tr>
<td>4.3.81 (ii)</td>
<td>3.0</td>
<td>16.0 × 3.5</td>
<td>15.8 ± 10.2</td>
<td>2.1</td>
<td></td>
<td>(2×6)</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
In all three earthquakes the surface slip predicted by the seismic source models (7-50 cm) is smaller than that observed (50-70 cm). This result is consistent with the proposal of Jackson et al. (1982) that a considerable amount of slip occurred on these faults following the seismic rupture. Papastamatiou & Mouyaris (1986) report evidence of aseismic slip for up to a week following the 1954 April 30 earthquake in central Greece so aseismic creep in the form of afterslip may not be uncommon. Aseismic afterslip is also consistent with the results of Billiris et al. (1991) who find that the seismic strain is only equal to about 60 per cent of

observed to the west of Pisia. The superior fit of the double source model to the waveforms of the March 4 earthquake implies that the offset between the two halves of the mapped surface break continues, as some type of unslipped barrier, to depth, and that two segments ruptured during this earthquake.

Figure 8. P-wave models (dashed lines) of displacement data (solid lines) for the 1981 February 25 earthquake (see Table 4). Amplitudes (nm) of the models are scaled to fit at each station using stress drop (Δσ), stations are at azimuth φ and epicentral distance Δ. Note that all the seismograms shown in this paper have negative first motions and the processing generated precursor of opposite polarity shown by some should be ignored. Moment-rate function as in Fig. 7.

Figure 9. 1981 March 4: preferred double source model (dashed lines) of the displacement P-wave seismograms (solid lines) (see Table 4). Annotation and moment-rate function as in Fig. 7.
Table 5. Comparison of the source parameters of the Gulf of Corinth (1981) earthquakes determined in this study with previous studies. JJ refers to Jackson et al. (1982), Kim to Kim et al. (1984), Bz to Bezzeghoud et al. (1986), and the CMT solutions are from Dziewonski & Woodhouse (1983). Note that the CMT depths are not well constrained for shallow earthquakes.

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Depth (km)</td>
<td>Moment (x10^18 Nm)</td>
<td>Δσ (bars)</td>
</tr>
<tr>
<td>This study</td>
<td>5.0</td>
<td>6.0</td>
<td>22</td>
</tr>
<tr>
<td>Jackson</td>
<td>10.0</td>
<td>7.3</td>
<td>34</td>
</tr>
<tr>
<td>Kim</td>
<td>12.0</td>
<td>8.1</td>
<td>10</td>
</tr>
<tr>
<td>Bz</td>
<td>6.0</td>
<td>10.0</td>
<td>-</td>
</tr>
<tr>
<td>CMT</td>
<td>20.0</td>
<td>12.9</td>
<td>-</td>
</tr>
</tbody>
</table>

the geodetic strain across central Greece over the last hundred years.

6 CONCLUSIONS

(1) Source parameters for the three normal faulting earthquakes, the largest events in the 1981 Gulf of Corinth sequence, are determined using broad-band P-wave modelling. The earthquakes are shallow, confirming that the seismogenic zone of the crust beneath Greece and the Aegean Sea is only about 10 km thick. They also all have low stress drops (<30 bars) and rupture velocities which may be explained by the low friction expected on normal faults in an extensional regime.

(2) The 2-D seismic source models of the Gulf of Corinth earthquakes are able to produce good fits to the waveforms and surface faulting compatible with that observed. They imply that the February 24 earthquake was responsible for

Figure 10. Map of the eastern end of the Gulf of Corinth (as in Fig. 6) showing the epicentres of the three largest 1981 earthquakes, and the two initial subevents (I and II) to the February 24 mainshock (M). Also shown are the surface breaks predicted by the preferred source models determined in this study, and the shaded areas represent the map view of the rupture areas. The February 24 earthquake model surface break clearly extends onshore on to the Perahora peninsula, its strike corresponding to the strike of the mapped faulting to the west of Pisia. The February 25 earthquake model surface break has the same strike as the mapped faulting to the east of Pisia, and its relative position is well within the location and depth errors of the source (approximately ±5 km, Jackson et al. 1982).
the surface faulting on the Perahora peninsula to the west of Psia, and that the February 25 earthquake produced the surface break to the east of Psia. The March 4 earthquake was a complex, probably double, event. The segmented pattern of surface faulting appears to continue to depth. (3) Small subevents (<1 per cent of the total moment) preceded the main energy release of all three Gulf of Cortin earthquakes by about 1 s. Such initiation of damaging earthquakes by small (M4–5) subevents is apparently not uncommon and implies that dynamic stresses may be as important as long-term static ones in controlling earthquake nucleation.

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REFERENCES


Source modelling of Aegean earthquakes


