Complexity of rupture in large strike-slip earthquakes in Turkey

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(Received October 1, 1981; accepted for publication October 28, 1981)


Complexity of rupture propagation has an important bearing on the state of stress along the earthquake fault plane and on the prediction of strong ground motion in the near-field. By studying far-field body waveforms recorded by WWSSN long-period seismograms it has been possible to investigate the degree of complexity of several Turkish earthquakes. The results, which are obtained by matching synthetic P waveforms to observed data indicate that the July 22, 1967 Mudurnu Valley earthquake (\(M_s = 7.1\)) is a complex event which can be explained by the superposition of elementary sources with variable amplitudes and source time sequence history. In this regard, it is very similar to the February 4, 1976 Guatemala earthquake (\(M_s = 7.5\)). A comparison of these two events indicates that their source-time series ranges from 5 to ca. 20 s and, regardless of the total moment of the earthquake, the moment of the individual events is bounded at around \(5 \times 10^{26}\) dyn cm. The November 24, 1976 E. Turkey earthquake (\(M_s = 7.3\)), on the other hand, has a complexity which cannot be explained by such a simple model; in this respect, it may be more similar to the Tangshan, China, earthquake and as such, may involve significant thrust, normal or other complications to its faulting mechanism than the strike-slip mechanism of the P-wave first-motion data. The source time history for the 1967 Mudurnu Valley event is used to illustrate its significance in modeling strong ground motion in the near field. The complex source-time series of the 1967 event predicts greater amplitudes (2.5 larger) in strong ground motion than a uniform model scaled to the same size for a station 20 km from the fault. Such complexity is clearly important in understanding what strong ground motion to expect in the near-field of these and other continental strike-slip faults such as the San Andreas.

1. Introduction

The tectonics of Turkey are dominated by an E–W-trending fault, the Anatolian fault (Fig. 1). It extends from western Turkey, south of the Sea of Marmara across the country to the east to its junction with the E. Anatolian fault. It is presumed to extend further eastward towards Van Gölü (Lake Van). A segment of the fault is presumed to lie to the northeast of the lake as shown in Fig. 1. The exact location is best determined in the central part of the country (N. Anatolian fault) while its position both to the west and east is less certain. Ambraseys (1970) has described its surface expression in detail.

Throughout historic times the Anatolian fault has been the site of many large and destructive earthquakes. Such events have been well-documented by Ambraseys (1971) from his investigations of Turkish historic records. Some events date back as far as 10 A.D.

Because of this extensive historic record and the occurrence of many large events this century, the Anatolian fault is an ideal locality for investigating earthquake prediction (Allen, 1982) and attempting to understand the rupture process in large strike-slip events. Additionally, being in a continental crustal environment, it is important to study large earthquakes occurring along the Anatolian fault as analogues for a great earth-
quake on the San Andreas fault in California.

Since the installation of the WWSSN network in 1963, several important events have occurred in Turkey. Especially significant are the 1967 Mudurnu Valley and 1976 E. Turkey events. Both were well-recorded by the WWSSN network as well as many other smaller events.

In this study four events occurring since 1963 are investigated. They are shown in Fig. 1 along with their corresponding focal mechanisms, the details of which are shown in Fig. 2. Event parameters are given in Table I. The events in 1964 ($M_s = 6.5$) and 1966 ($M_s = 6.8$) are smaller than, and are used for comparison with, the events of major interest here—the 1967 ($M_s = 7.1$) and 1976 ($M_s = 7.3$) earthquakes.

### TABLE I

<table>
<thead>
<tr>
<th>Date</th>
<th>Origin time (U.T.) (h min s)</th>
<th>Location $^\circ$N $^\circ$E</th>
<th>$M_s$ $^a$</th>
</tr>
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<tr>
<td>Turkey</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>October 6, 1964</td>
<td>14 31 23.0</td>
<td>40.30 28.23</td>
<td>6.5</td>
</tr>
<tr>
<td>August 19, 1966</td>
<td>12 22 10.5</td>
<td>39.17 41.56</td>
<td>6.8</td>
</tr>
<tr>
<td>July 22, 1967</td>
<td>16 56 58.0</td>
<td>40.67 30.69</td>
<td>7.1</td>
</tr>
<tr>
<td>November 24, 1976</td>
<td>12 22 16.0</td>
<td>39.05 44.04</td>
<td>7.3</td>
</tr>
<tr>
<td>Tangshan</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>July 27, 1976</td>
<td>19 42 54.0</td>
<td>39.56 117.87</td>
<td>7.7</td>
</tr>
<tr>
<td>Guatemala</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>February 4, 1976</td>
<td>09 01 43.9</td>
<td>15.28 89.19</td>
<td>7.5</td>
</tr>
</tbody>
</table>

$^a M_s$ values are determined by the National Earthquake Information Service (NEIS) of the U.S. Geological Survey.
Fig. 2. P-wave first-motion data and focal mechanisms for the events shown in Fig. 1 and listed in Table I. Events 2, 3 and 4 are well constrained by the first-motion data. The sense of faulting is indicated on the presumed fault planes.
2. Body wave analysis

2.1. Introduction

Although it is important to understand the source rupture process in simple events, a greater challenge lies in attempting to understand the rupture process when more than one source is involved.

Previous studies have attempted to unravel such complexity. Take, for example, the 1971 San Fernando earthquake. Langston (1978) and Heaton (1982) have both attempted to understand the details of this small- to intermediate-size earthquake. Both found it to be quite complex and have proposed rupture on two faults of different orientation to explain the observed seismological data. Undoubtedly, with the number of unconstrained parameters for this event, it is almost certainly possible to find additional models that are just as compatible with the observed data.

Unless one can constrain the depths and mechanisms of the multiple sources well, the unknown parameters are too numerous for a predominantly thrust- or normal-fault source sequence to be well constrained.

On the other hand, some simplifying assumptions can be introduced in dealing with the complexity associated with rupture propagation in large strike-slip earthquakes. In these events the depth of faulting is usually restricted to the upper 20 km. Since rupture takes place along the strike of such faults, one can assume that such rupture consists of several point sources, located, for instance, at 10 km depth, distributed along the fault at varying intervals.

Because of the geometry of strike-slip faults (ribbon-like), it is possible to assume that each source has an almost identical mechanism, since the change in the strike direction for such faults is usually not very significant (<20–30°). Other variables which are of interest are the source-time function and finiteness over the total rupture plane. These are important, and may have to be included if the nature of the rupture appears to demand them.

2.2. Events of 1964 and 1966

To demonstrate that the complexity exhibited by the 1967 and 1976 events is indeed real and not due to propagation or near-receiver effects, a smaller event near each of the complex events was selected for comparison. The October 6, 1964 and August 19, 1966 events were chosen as being suitable events close to the 1967 and 1976 events, respectively. Their locations are shown in Fig. 1 with the details of their focal mechanisms illustrated in Fig. 2. The 1964 event is a predominantly normal fault event. The fault planes, however, are not well constrained. McKenzie (1978) gives a similar solution for this event. The solution obtained for the 1966 event agrees well with that proposed by McKenzie (1969) but disagrees with the almost pure strike-slip solution obtained by Nowroozi (1972). Further details of the faulting and damage effects of the 1966 event can be found in Ambraseys and Zátopek (1968) and Wallace (1968).

The relative simplicity of the 1964 and 1966 events compared with the complex 1967 and 1976 events can be seen in Fig. 3. This figure shows representative P-wave seismograms for all the Turkish events studied here as well as for the February 4, 1976 Guatemala earthquake studied in detail by Kanamori and Stewart (1978) and the July 27, 1976 Tangshan, China, earthquake studied by Butler et al. (1978). The event parameters are given in Table 1. Clearly compared with the others, the 1964 and 1966 events are simple and most probably consist of a single source.

As regards the complex events, each exhibits some characteristic differences. The 1976 event, for example, has a higher frequency content than the 1967 event. Also the largest trace amplitudes occur towards the end of the trace. Clearly, for the 1967 event the largest pulse is at the start. The characteristic period of the Guatemala trace appears to be longer than any of the others, as is its total duration. As with the 1976 event, the larger amplitudes are to be found towards the later part of the record. The Tangshan event appears to be much shorter in duration with most of the energy arriving in the first minute of record.

Examining only one seismogram for a particu-
lar earthquake may result in an incorrect interpretation for the event. To avoid this, and to gain a better overall view of the characteristics of the source, seismograms were selected for examination from as many azimuths and suitable distances (Δ = 30°–90°) as possible. This procedure was adopted for all of the events discussed here. The simplicity suggested by the WES P waveform for the 1964 event seems to be representative of the waveforms at many other azimuths as indicated in Fig. 4. The only station that shows any complexity is Ponta Delgada, Azores (PDA), and this is probably due to its noisy island location. The P-wave seismograms for the 1966 event, the other simple earthquake, are shown in Fig. 5. The record at College, Alaska (COL), and the waveforms on the left-hand side of the figure are simple, for the most part, although not as simple as those of the 1964 event. On the right-hand side of the figure the waveforms appear somewhat more complex. However, this is probably due to their being located close to a double node so that the background noise is amplified relative to the low signal for these stations. The principal differences between these two events is most likely the result of their different mechanisms. Pure 45° thrust or normal events give rise to symmetric radiation patterns with regard to P-wave excitation as a function of azimuth in the far-field. The 1966 event is clearly more complex than this. However, both events show simple enough waveforms to be used for comparison purposes. Thus the greater complexity demonstrated by the 1967 and 1976 events is due to a source effect and not to a propagation or near-receiver effect.

### 2.3. Guatemala event

The main aim of this study is to investigate the complexity of the July 22, 1967 Mudurnu Valley and November 24, 1976 E. Turkey events. Both these events have almost pure strike-slip mechanisms as shown earlier. The method of analysis is the same as that discussed by Kanamori and Stewart (1978) for the February 4, 1976 Guatemala earthquake. The results of their P-waveform modeling are shown in Fig. 6. For the Guatemala event many stations in Europe remained on-scale and were usable. The only other usable station was La Paz, Bolivia (LPB), and because of its close distance only the portion of the record prior to the arrival of PP was modeled. A source time function of 9 s duration was used in the modeling procedure with the direct P and surface reflected pP and sP included. The function shown in Fig. 6 was minimized and the resulting source time series and synthetics obtained. The data quality and distribution did not warrant a more detailed approach such as including finiteness in the source or changing the mechanism for the different sources. A total of approximately ten sources appear to be
present in the source sequence. The source time series for all the stations are replotted in Fig. 7. Events 1, 2, 3, 6, 7, 8 and 9 are quite clear, the others less so. One of the drawbacks of this study was the lack of good azimuthal control. All the European stations lie within 23° of each other and only one other azimuth (LPB) is represented. Note from this figure that events 8 and 9 appear to be the largest in the sequence. They occur 70–100 s after the initial arrival.

2.4. Event of 1967

The Mudurnu Valley earthquake of July 22, 1967 (\(M_e = 7.1\)) occurred along the N. Anatolian fault. Ambraseys and Zátopek (1969) reported the effects of damage and faulting caused by the earthquake. Their study indicated right-lateral faulting over a distance of 80 km. Although significant, this value is smaller than several other large events in the region. For example, the December 26, 1939 event (\(M_e = 8.0\)) had 350 km of rupture length associated with it, while the events of November 26, 1943 (\(M_e = 7.6\)) and February 1, 1944 (\(M_e = 7.6\)) had 270 and 190 km, respectively. In all cases, the displacement was predominantly right-lateral.

Nevertheless, the 1967 event represents the largest earthquake along the N. Anatolian fault to be recorded by the WWSSN since its operation began in 1963. Consequently, an analysis of such data would contribute to a better understanding of the mechanism of rupture along this important tectonic boundary. Such an analysis would help to
Fig. 5. Vertical long-period seismograms for various WWSSN stations in the distance range 30°-90° as a function of azimuth around the epicenter of the 1966 event. Note the fairly simple P waveforms at most stations. All stations are plotted with the same vertical scale. The stations to the east are close to a double node and so are of small amplitude. The apparently more complex looking waveforms are due to an amplification of the noise relative to the signal for these azimuths. The open and closed circles indicate the locations on the focal sphere of the stations shown. The fault planes are constrained by the data shown in Fig. 2.

predict the character of future activity in this region as well as in other regions of major continental strike-slip faults such as the San Andreas.

The P-wave first-motion study of the 1967 event indicates an almost pure strike-slip solution (Fig. 2), with right-lateral motion along the east-west fault plane. Since the P-wave coda indicated a complex wavetrain (Fig. 3), the technique devised by Kanamori and Stewart (1978), in their study of the Guatemala earthquake, was applied in the hope of understanding better the nature of the complexity. Normally, for such a large event one would expect the P waves recorded on the WWSSN to be off-scale. However, due to the orientation of the P-wave radiation pattern from a vertical strike-slip fault the energy in the far-field is small enough to allow for P waves to be well-recorded on many of the WWSSN seismograms. The poor azimuthal control found in the Guatemala study is not a problem here. Ten good-quality records are used with good azimuthal coverage (Fig. 8).

After some preliminary analysis a time function of 10 s duration was selected for use in the modeling process. A P-wave synthetic $s(t)$ was calcu-
Fig. 6. Observed and synthetic P waveforms for individual WWSSN stations from the multiple-shock analysis. For each station the source time series is obtained by using the mechanism of the mainshock of the Guatemala earthquake and the source time function shown here. The surface reflections pP and sP are included in the source time function. The resulting series is given for each station along with the moment for the first event. The height of the vertical bar is proportional to the moment of the individual event; Δ is the epicentral distance. This figure is taken from Kanamori and Stewart (1978).
lated for each of the ten stations assuming the mechanism shown in Fig. 2 and a depth of 7 km. As shown in Fig. 8, the synthetic consisted of the sum of the direct P and the surface reflections pP and sP. The synthetic pulse was added in a time series changing only the amplitude and time interval between pulses until the whole P-wave record was modeled for each station. The best fit to the observed record was found by a non-linear least-squares fit. The results for all the stations modeled are shown in Fig. 8. The source time series below each observed and synthetic indicates the number, amplitude and time spacing of the sources used to obtain the above synthetic record. The fits obtained are good considering the simplified method used and that the same depth, mechanism and time function is used for each of the sources. The most notable differences between the observed and synthetic records occurs during the initial cycle. The higher frequency nature of the first pulse compared with the others obviously requires a shorter-duration time function for its exact modeling. Such a time function would be inappropriate for the modeling of the later arrivals, however. To simplify the procedure the time function of 10 s duration was kept throughout. The resulting source time series obtained for each station are replotted in Fig. 9. Note that although there are differences from station to station the overall agreement is good. The shaded sections represent the range of times in which individual sources occur. The data indicates seven sources occurring in the first 2 min of record.

The largest source in each case is the first event. On the right-hand side of the figure the moment of this event is given together with the total moment of all sources. The average of the total moment is $1.5 \times 10^{27}$ dyn cm. This value is approximately twice that reported by Hanks and Wyss (1972). However, they used only part of the record in their analysis of body-wave spectra.

2.5. Event of 1976

The November 24, 1976 E. Turkey earthquake ($M_s = 7.3$) occurred along a section of fault northeast of Van Göllü. The event parameters are given in Table I. As shown in Fig. 1 the epicenter of the 1976 event lies close to the border of Turkey with Iran. Toksöz et al. (1977) reported that this was the only large earthquake to occur during the last century in this area according to a review they made of seismicity catalogues and the recollections of villagers. At the event’s epicenter the fault strikes east-southeast. The association between this portion of the fault and the extension of the N. Anatolian fault is not clear although it is presumed that the two are part of the same tectonic province.

Evidence from field studies (Toksöz et al., 1977) indicated 55 km of right-lateral strike-slip rupture associated with the 1976 event. The average observed displacement reported was 2.5 m. The observed field evidence of right-lateral strike-slip motion is confirmed by the P-wave first-motion plots.
Fig. 8. Observed and synthetic P waveforms for individual WWSSN stations obtained from the multiple-shock analysis. For each station the source time series is obtained by using the mechanism shown in Fig. 2 (No. 3) and the source-time function shown here. The surface reflections pP and sP are included in the source time function. The resulting series is given for each station along with the moment for the first event. The height of the vertical bar is proportional to the moment of the individual event; \( \Delta \) is the epicentral distance and \( \phi_s \) the station azimuth.

shown in Figs. 1 and 2 and Toksöz et al. (1977).

As indicated in Fig. 10, P-wave seismograms show a complex wavetrain. Compared with the other events shown in Fig. 3, the 1976 event has a high frequency content overall, with most of the energy arriving in \( \sim 1.5 \) min. As with the Guatemala event, the higher amplitude energy arrives later in the sequence. This is to be contrasted with the 1967 event where the largest energy arrives in the first pulse.

Following the procedure discussed earlier for the Guatemala and Mudurnu Valley events, an
attempt was made to model the complex sequence of arrivals of the 1976 event. Again, due to its location, a good azimuthal station coverage was obtained. The nine WWSSN stations selected for study are shown in Fig. 10. In this case, a shorter time function was selected (4.5 s duration). The resulting best-fitting synthetics are shown below the observed data. Also shown are the corresponding source time series.

The impression on visually comparing the observed with the synthetic data is that the fits are poor, at least not as good as those found for the Guatemala and Mudurnu Valley events. This becomes clearer when the source time series are replotted vertically as in Fig. 11. The variability in amplitude and time separation from event to event is considerable. Events 1 and possibly 2 show consistency from station to station. For all of the others, however, the variability is unacceptable. Several different time functions and depths were tried but what is shown represents the best fits that could be obtained.

From this analysis it can be concluded that not all large strike-slip events can be fitted using this method. There are several reasons for this. In the modeling procedure, several assumptions were made about the individual sources. The depth was considered to be the same for all the events. Using the assumption of a shallow strike-slip source it seems unlikely that varying the depth from 0 to 20 km, for example, could result in enough variation in the synthetics to explain the observed waveforms. A variable time function for each source would similarly be unlikely to explain the complexity. The most likely explanation of poor fits is that the assumptions regarding the source mechanisms of the individual sources is inappropriate in this case. Since fault ruptures are rarely straight it is reasonable that the later events have a different mechanism from the mainshock; the assumption that was used in the modeling of the Guatemala, Mudurnu Valley and E. Turkey events was that the source mechanisms of the later events were identical to that of the mainshock. However, for this effect to be significant the mechanism should be considerably different. It seems unlikely that rotating a fault plane by a few degrees to take account of the fault curvature would be sufficient. However, if some of the later sources had mechanisms very different from the strike-slip mechanism of the main event, then it would be very difficult to unravel the details. Perhaps a thrust or normal fault event on an associated fault structure may have been triggered by the main event. Toksöz et al. (1977) in their field study of the earthquake made no mention of any such associated faulting. However, because of the nature of the terrain and the winter conditions prevailing at the time of the earthquake, such an observation could easily have gone unnoticed. Furthermore, such an event may not have caused fault rupture at the surface.
3. Discussion

In the above analysis an attempt has been made to understand the complexity of two Turkish earthquakes and compare the results with those obtained previously (Kanamori and Stewart, 1978) for the Guatemala earthquake. From a study of both Turkish events a satisfactory interpretation was found only for the Mudurnu Valley, 1967 event. At this time, it was not possible to obtain a complete and satisfactory interpretation for the E. Turkey, 1976 event.

The amplitude values averaged over all the WWSSN stations used in each analysis are shown in Fig. 12 for the Guatemalan and Mudurnu Valley events. The numerical values are shown at the top of Figs. 7 and 9, respectively. Figure 12 summarizes the results of this study. The events differ by a factor of 2 in seismic moment. However, the maximum moment of an individual event is bounded at about $5 \times 10^{26}$ dyn cm, regardless. Each of these complex events can be expressed as
a sequence of distinct events that occurred at 5 to approximately 20 s intervals as shown in Fig. 12. The average time separation is 11.1 and 17.5 s for the Guatemala and Mudurnu Valley events, respectively.

In a recent study, Kikuchi and Kanamori (1982) re-examined the complex P-wave data for the Guatemala earthquake. They found essentially the same result as Kanamori and Stewart (1978), using a single source time history to explain all the data. Understanding such complexity is clearly important in predicting what strong ground motion to expect in the near-field of these and other continental strike-slip faults such as the San Andreas. Haskell (1966) and Aki (1967) showed that irregular fault motion significantly enhances the high-frequency end of the seismic spectrum. This is shown in Fig. 13. The upper figure (random model) schematically represents what occurred during the 1967 Mudurnu Valley earthquake. The fault is 80 km long and it is assumed here that the rupture took place from right to left in a unilateral sense. The individual sources are shown as solid circles along the fault. Each source is scaled according to the results shown in Fig. 12. The sources are spaced equally along the fault and occur at the times indicated in Fig. 12. The solid square represents a station 20 km from the point of initial rupture. The long-period ground motion resulting from the fault rupture is computed at this station. The observed ground motion from the 1968 Borrego Mountain earthquake is used as a Green's function for each point source. The ground displacement is computed for the station by a superposition of the sources, appropriately scaled and time delayed. This method is discussed in greater detail by Kanamori (1979). The resulting horizontal vectorial displacement is shown. The maximum amplitude computed for the random model is 40 cm. The calculation is repeated, this time replacing the Mudurnu Valley source time series with a uniform model. The overall size (total
an attempt has been made to analyze two of the largest Turkish earthquakes recorded since the installation of the WWSSN network in 1963, i.e. the Mudurnu Valley, 1967 event and the E. Turkey, 1976 event. The complexity observed in the P-wave seismograms was shown to be the result of source effects since two smaller events close to these in 1964 and 1966 showed only simple waveforms in the far-field, ruling out the possibility of propagation or near-receiver effects.

The method of analysis followed that of Kanamori and Stewart (1978) for the Guatemala earthquake. The results for the 1967 event were quite successful in that the waveforms could be modeled as a series of individual sources of varying amplitude and time sequence history. Comparing the Guatemala and Mudurnu Valley data it can be said that, based on these events, the source-time series ranges from 5 to ca. 20 s and, regardless of the total moment of the earthquake, the moment of individual events is bounded at around $5 \times 10^{26}$ dyn cm.

On the other hand, the 1976 event analysis was not as successful. For this event it does not appear possible to model the event as a multiple source with each source having the same mechanism as that obtained from the P-wave first-motion data. A radical variation in the individual source mechanisms from that of the first event seems necessary, perhaps involving large components of thrust or normal faulting on associated neighboring faults.

The multiple-source character demonstrated by the Guatemala and Mudurnu Valley events has significance with regard to the prediction of strong ground motion. The strong ground motion was computed at a station 20 km from the Mudurnu Valley rupture zone assuming the model derived here. The resulting motion was 2.5 times greater than that obtained from a model of uniform displacement, all other parameters being kept the same.

An understanding of the mode of rupture in these events has important bearing with regard to other environments of continental strike-slip tectonics such as the San Andreas in California.

4. Conclusions

From an analysis of P-waveform data recorded by the WWSSN stations at teleseismic distances

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Fig. 13. Displacements at 20 km from the fault (solid square), calculated for the random model (the specific example shown is for the Mudurnu Valley, Turkey, event, shown in Fig. 12, assuming a unilateral propagation of the source) and the uniform model. The total moment in each case is the same, only the distribution of the energy is different. Note that the random model leads to larger displacements than the uniform model.

moment) is kept the same but now the sources are of equal size and are spaced equally along the fault. They occur also at equal time intervals. The resulting displacement is plotted below. The resulting amplitude is decreased by more than one half that estimated for the random model.

Thus, in estimating ground motion for a site near a fault, it is important to know whether the expected rupture will be uniform or occur as a series of random sources.
Acknowledgments

We acknowledge useful discussions with Nafi Toksöz in regard to the faulting process of the 1976 E. Turkey event.

This research was supported by a grant from the National Academy of Sciences, through WDC-A for seismology and the U.S. Geological Survey Contract No. 14-08-0001-19265. Contribution No. 3705 Division of Geological and Planetary Sciences, California Institute of Technology, Pasadena, CA 91125, U.S.A.

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