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GROUND MOTION WAVEFORMS AND SOURCE SPECTRAL SCALING FROM CLOSE-DISTANCE ACCELEROMETERS IN A COMPRESSIONAL REGIME AREA (FRIULI, NORTH-EASTERN ITALY)

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ABSTRACT

An aftershock sequence was recorded by an SMA-1 instrument installed at Gemona del Friuli in a massive limestone site during the period from May to October 1976. The sequence consists of small to moderately strong, dominantly thrust-faulting earthquakes occurring at close distance \( R \approx 10 \text{ km} \) from the station. For twenty-four of them, the instrument was triggered by \( P \)-waves; the resulting records appear to be particularly suitable to the investigation of source properties for the Friuli earthquakes. After the digitization of the film traces by means of an automatic high resolution (\( \sim 400 \text{ sps} \)) optical scanner, a simple numerical technique has been applied to obtain ground-motion displacement directly from digitized accelerograms by deconvolving the instrumental response in the time domain. The resulting acceleration, velocity and displacement waveforms have been analyzed in order to investigate source properties. Durations \( T \) and areas \( \Omega_s \) of the shear-wave pulses have been measured on the displacement time histories and compared with the values of corner frequency \( f_c \) and seismic moment \( M_0 \) estimated in the frequency domain from the acceleration spectra of the same events. After correcting for attenuation, the scaling of source parameters indicates that stress drops are approximately constant \( (\Delta \sigma = 300 \text{ bars, on the average}) \) in the seismic moment range from \( 10^{21} \) through \( 10^{25} \) dyne-cm.

For this sequence, the behavior of peak ground motions versus earthquake size is consistent with the trend obtained from stochastic simulations assuming an omega-square spectral model with a constant stress drop of 300 bars. Moreover, the availability of objective waveforms of ground displacement in a wide seismic moment range offers a favourable opportunity to study the scaling of the ground displacement pulse. We find that the apparent breakdown of similarity depicted by the quantity \( \frac{d_p}{M_0} \) (\( d_p \) is ground displacement peak) in the low seismic moment range \( (M_0 \leq 10^{23} \text{ dyne-cm}) \) is more a dissipation effect than a source property: after correction for attenuation, the scaling of \( \frac{d_p}{M_0} \) versus seismic moment agrees with a constant stress value of 300 bars. The results obtained in this paper enhance the bias that dissipation mechanisms can produce in the interpretation of source properties, even at close focal distances.

Finally, the applicability to Friuli earthquakes of an omega-square model with a constant stress drop of 300 bars is further confirmed by an analysis of spectral ratios between different size earthquakes of the sequence: this analysis demonstrates that the finding of this spectral scaling model is not conditioned, to first order, by whole-path or near-site propagation properties whose effect can be significant when data from a single station are investigated.
INTRODUCTION

On May 6th, 1976, at 20:00 GMT a strong earthquake struck the Friuli region, northeastern Italy, producing extensive damage and more than 1,000 deaths. Its average local magnitude was $M_L = 6.5$, very close to the value 6.6 obtained from Wood-Anderson records synthesized using strong-motion accelerograms (see Bonamassa and Rovelli, 1986). Within a day, several national and international seismological institutions converged on the epicentral area, installing networks of short-period seismometers and accelerometers. Among these institutions were the Ente Nazionale per le Energie Alternative (ENEA), the Ente Nazionale per l'Energia Elettrica (ENEL), and the Istituto Nazionale di Geofisica (ING). While the efforts of the ENEA-ENEL Joint Commission were mainly devoted to the collection of strong-motion data, the principal interest of the ING was recording microearthquakes. However, the ING installed one SMA-1 accelerometer in the most convenient site among those selected for the seismometric stations. This accelerometer was operated at Tarcento (from 8 up to 9 May) and at Somplago (from 9 up to 22 May). Afterward, the instrument was moved to Gemona del Friuli (from 22 May up to the end of October 1976), only a few kilometers from the mainshock epicenter (see Figure 1).

Figure 1 - Main neotectonic structural features, epicenter distribution and available fault plane solutions for some aftershocks recorded at Gemona del Friuli. Open circles represent the location of the four largest aftershocks; full circle points out the mainshock of May 6th, 1976. Full square gives the location of the Gemona station. The fault plane solutions are taken from the literature (Cipar, 1980; Anderson, 1985; Giardini et al., 1985; Barbano et al., 1985; Renner and Slejko, 1986) and enhance a compressional regime characterized by thrust or reverse mechanisms. 1: dip-slip fault (dashes towards lowered area, arrows according to the dip direction); 2: strike-slip fault; 3: fault with undetermined character; 4: uplifting axis; 5: anticline axis; 6: asymmetrical uplift (arrow towards less uplifted area). Redrawn from Barbano et al. (1985).
The Friuli sequence was characterized by a large number of aftershocks from small to moderate magnitude. In particular, starting from September 9th, 1976, a significant increase of seismic activity took place, which produced two earthquakes with magnitude comparable to that of the mainshock (both occurred September 15th at 03:15 and 09:21 GMT, respectively). The accelerometer installed at Gemona del Friuli allowed the collection of a large number of strong-motion recordings from events in a wide magnitude range at close focal distance. Of all the accelerometer locations in Friuli, Gemona was the nearest one to the epicenters.

The Friuli sequence of 1976 has been studied in some previous papers (e.g. Cipar, 1980 and 1981; Lyon-Caen, 1980; Barbano et al., 1985). In these studies teleseismic, regional and local data were used to investigate the seismogenic features of the active faults in that area. Source time functions for the three largest events of the sequence (May 6th at 20:00 GMT, September 15th at 03:15 and 09:21 GMT, respectively) were investigated by Cipar (1980 and 1981), who estimated seismic moment and source duration using teleseismic and regional data. Very similar results were found by Lyon-Caen (1980) who relocated the events, computed focal mechanisms and analyzed the body waves of short- and long-periods. Moreover, abundant strong-motion accelerograms were recorded during the 1976 sequence by the ENEA-ENEL network. In particular, these data were used in two recent papers (De Natale et al., 1987; Cocco and Rovelli, 1989) in order to study the high-frequency behavior of seismic radiation and source spectral scaling.

The availability of further close-distance accelerograms, recorded at the same station for a wide magnitude interval provides a further occasion to investigate the characteristics of the seismic source for the Friuli earthquakes.

The uniqueness of the collected data set, although recorded by a conventional analog instrument, demands an effort to extract as much information as possible about ground motion properties at close distance, including waveforms of ground velocity and displacement. To do this, a particular attention had to be given to high-resolution digitization and an appropriate algorithm to improve the quality of the corrected time histories of ground motion particularly for weak events (often recorded with very unfavourable signal-to-noise ratios). This paper aims to show some results obtained from these data.

**The Data Base**

During the period of observation at Gemona del Friuli, more than fifty recordings of aftershocks were collected, 24 of which were triggered by P-waves. In the latter case, the entire S-wave train is contained in the accelerogram, allowing analyses devoted to determine source and/or propagation properties. We chose these 24 events as the data base for this study.

The instrument used at Gemona del Friuli was a Kinemetrics SMA-1. It recorded photooptically the three components of the ground motion on 70mm film, with 1 g full scale range. On both the top and bottom of the film, two traces display the time marks provided by an internal (not absolute) timer. The accelerograph was installed on the basement of a two-story building founded on massive Mesozoic limestones. Its orientation was such that the longitudinal axis pointed N07W.

The lack of absolute timing on the recordings might cause some difficulties in the identification of the single events. The occurrence of many aftershocks recorded on the same film, however, allowed us to obtain a reliable association between our recordings and the events re-
ported on the seismological bulletins, at least for the most significant earthquakes. The resulting data set is listed in chronological order in Table 1.

In order to obtain good quality ground-motion waveforms, the digitization technique deserves particular attention: this operation has been carried out by means of a high-resolution optical scanner, giving 162 counts per inch corresponding to a sampling interval $\Delta t$ equal to 0.00244 sec in the time axis. Another relevant feature that required careful attention was the instrumental deconvolution and the estimation of the ground displacement from the accelerograms. We have adopted a procedure that yields ground displacement by deconvolving digitized accelerograms in time domain. The deconvolution operator has been analytically derived from the Laplace transform of the differential equation describing the recording system when the input is given by a Dirac function. This is described in detail in the paper by Alessandrin 

THE FRIULI SEQUENCE: SEISMOTECTONIC FEATURES

The hypocentral distribution and the focal mechanisms of the largest Friuli shocks in the years 1976-1979 have been discussed by Barbano et al. (1985). They relocated these earthquakes and related them to the Dinaric structures involved in the south-Alpine shortening. The main neotectonic structural features have been drawn in Figure 1: the most important structural evidences are the E-W-striking south-Alpine overthrusts and the SE-NW-striking Dinaric overthrusts. It is interesting to observe that the mainshock and the stronger aftershocks occurred on the E-W-striking south-Alpine system.

Figure 1 shows the epicenter location of the four largest aftershocks of the sequence; the epicenters have been taken from Barbano et al. (1985). We avoided to include in this figure the epicenters of the other smaller events of the Gemona data set because of their larger uncertainty. However, taking into account that the difference between $S$-wave arrivals and triggering time (all the accelerograms analyzed in this paper were triggered by $P$-waves) gives a minimum distance value, we may infer that focal distances for our data set range between 7 to 15 km. For the largest events reliable depth estimates are available (Zonno and Kind, 1984; Barbano et al., 1985), revealing the shallow depth of the Friuli earthquakes: their values are in the range between 5 to 10 km.

The fault plane solutions were estimated by many authors for the four largest aftershocks (event 5, 6, 13 and 17 of Table 1) using waveforms or polarities at teleseismic distance (Cipar, 1980; Anderson, 1985; Giardini et al., 1985). For seven other aftershocks of the data set (event 1, 3, 10, 11, 15, 18 and 21 of Table 1), fault mechanisms were recently computed using data from local networks (Barbano et al., 1985; Renner and Slejko, 1986) based on polarities of first arrivals. Figure 1 shows all the available focal mechanisms. Fault-plane solutions of Figure 1 indicate a compressional regime characterized by thrust or reverse mechanisms. Unfortunately, focal mechanisms were not available in the literature for the other events included in Table 1 because of the paucity of both information about polarities of seismograms in the bulletins and waveforms at teleseismic distances with a satisfactory signal-to-noise ratio. However, the analysis of the polarization vector obtained from the horizontal components of the ground motion recorded at Gemona del Friuli reveals that, for more than 70% of the events, a stable wave polarization
characterizes both small and large aftershocks. This suggests that the smallest events of this sequence probably had thrust or reverse faulting mechanisms.

**Analysis of Ground Motion in Time and Frequency Domain**

The ground motion parameters have been already investigated by De Natale *et al.* (1987) and Cocco and Rovelli (1989). In particular, De Natale *et al.* (1987) estimated the seismic moments and corner frequencies applying an inversion procedure to the spectra of strong-motion data recorded in Friuli by the ENEA-ENEL Joint Commission. Cocco and Rovelli (1989) investigated the same strong-motion data used by De Natale *et al.* (1987), but restricted their analysis only to those events (seven) for which seismic moment could be supported by information available from data at teleseismic distance. Their estimates of stress drop exhibit a smaller dispersion compared to those computed by De Natale *et al.* (1987), Brune stress drop ranging between 250 and 400 bars for the largest Friuli aftershocks.

The availability of further accelerograms from other aftershocks in that area allows to reanalyze the scaling laws for the Friuli earthquakes. The good quality obtained for velocity and displacement time histories (see Alessandrini *et al.*, 1989) suggested to estimate source parameters both in time and frequency domain.

In the time domain, the joint inspection of the acceleration, velocity and displacement time histories on both the horizontal components allows the identification of source duration $T$, and the area of the displacement pulse $\Omega_o$ yields estimates of seismic moment $M_o$. The problems related to the duration estimates have been largely discussed by Boatwright (1984). In practice, source durations $T$ were estimated looking at the displacement and velocity waveforms simultaneously, and comparing them also with the squared velocity plots. Figures 2 and 3 show some examples of time histories of ground motions collected at Gemona del Friuli; in particular, the shaded areas indicate the duration estimated for those earthquakes. In these figures we can observe that the maximum amplitude phases of the acceleration time histories are distributed over the time window corresponding to the estimated source duration. Both Figures 2 and 3 ($a$ and $b$) reveal that the identification of the source duration is quite simple for the smaller earthquakes of the sequence, while difficulties increase for the largest events where incoherent arrivals characterize the accelerogram (Figures 2 and 3, $c$ and $d$).

Following Brune (1970), we have estimated seismic moment $M_o$ from the area of the displacement pulse $\Omega_o$ by using the well known equation (see also Frankel, 1981)

$$M_o = \frac{R \Omega_o}{C}$$

where $R$ is hypocentral distance and $C$ is a constant given by

$$C = \frac{F_s R_{\delta,\varphi} P}{4\pi\rho\beta^3}$$

including the effects of the free surface correction ($F_s = 2$), the rms average radiation pattern ($R_{\delta,\varphi} = 0.63$) and the partition of energy between the two horizontal components ($P = \frac{1}{2}$); $\rho$ and $\beta$ are density and velocity in the lithosphere (the values 2.7 $g/cm^3$ and 3.2 $km/sec$ have been used, respectively). Time-domain estimates of seismic moment and source duration for the
FIGURE 2 - Specimens of corrected ground motion waveforms derived from accelerograms recorded at Gemona del Friuli. A simple shape characterizes the small magnitude earthquakes (a and b, event 16 and 1 of Table 1, respectively), while complexity is peculiar to the strongest earthquakes of the sequence (c and d, event 17 and 5, respectively).
FIGURE 3 - Specimens of corrected ground motion waveforms derived from accelerograms recorded at Gemona del Friuli. A simple shape characterizes the small magnitude earthquakes (a and b, event 12 and 20 of Table 1, respectively), while complexity is peculiar to the strongest earthquakes of the sequence (c and d, event 6 and 13, respectively).
aftershocks recorded at Gemona del Friuli are listed in Table 1.

In our computations, the impedance variation between source and station site has not been taken into account. However, the Gemona station is situated on Mesozoic massive limestone, and tomographic studies for Friuli (Amato et al., 1989) indicate a small velocity gradient in the area around Gemona. Consequently, we have excluded that impedance variation could significantly affect our estimates of source parameters. This assumption is further discussed in the following.

The measurement of displacement pulse area might have been complicated by the source complexity, especially for the largest events (see Figures 2 and 3, c and d). In particular, for complex earthquakes the estimation of source parameters from strong motion data recorded in the near-source region may be primarily influenced by the size of the predominant sub-event.

The time-domain analysis has been intentionally carried out without any correction of the observed waveforms for path dissipation, assuming that attenuation effects other than geometrical spreading can be neglected owing to the small focus distance. In the following section, we will discuss to what extent this assumption is correct.

As far as the frequency-domain analysis is concerned, several techniques could be applied to the data in order to estimate source parameters. Our experience (Rovelli et al., 1988 a) suggests that visual inspection or the simultaneous inversion of the ground-motion spectrum in terms of both seismic moment $M_0$ and corner frequency $f_c$ have to be avoided. This because of the large uncertainty in the former approach, and because of the difficulties in controlling the bias on the simultaneous estimation of the two model parameters, due to their analytical dependence, in the latter approach. The investigation of the low-frequency behavior of the spectrum is particularly difficult in the analysis of strong-motion data, because of the lack of information peculiar to these data in the low-frequency band. In order to overcome this limitation, Rovelli et al. (1988 a) and Cocco and Rovelli (1989) used the information derived from independent data (seismograms at teleseismic distance or geodetic measurements, for instance, particularly suitable to infer large-scale features of the fault slip) to reduce the number of the unknown source parameters. Those authors adopted the values of seismic moment obtained from the literature, and used strong-motion data to compute the only unknown remaining parameter ($f_c$) of the source spectral model. This is justified by the fact that $M_0$ is controlled by the large-scale source properties, while corner frequency or stress drop are measures of the high-frequency ground motion, and strong-motion data are particularly well suited to investigate the high-frequency band.

Unfortunately, among the 24 aftershocks analyzed in this study, the seismic moment at teleseismic distance is available only for the two largest events. For this reason we were forced to estimate $M_0$ from the Gemona accelerograms. As we preferred to avoid the two-parameter inversion of the ground motion spectrum, we have primarily computed the corner frequency by means of an objective (moment independent) technique and afterward we have estimated the seismic moment from the ground motion spectrum. Following Rovelli et al. (1988 a) and Cocco and Rovelli (1989) we have applied the technique proposed by Andrews (1986) to determine corner frequency by means of the formula

$$ f_c = \frac{1}{2\pi} \frac{\int_0^\infty V^2(f)df}{\int_0^\infty D^2(f)df} \quad (2) $$
where $V(f)$ and $D(f)$ are velocity and displacement spectra, respectively. $V(f)$ and $D(f)$ have been computed by means of the following expression

$$
R \cdot \left( \frac{A(f)}{(2\pi f)^n} \right) e^{\pi \kappa f} e^{-\frac{\pi f R}{Q_o}}
$$

using $n = 1$ and $n = 2$, respectively. $A(f)$ is the spectrum of ground acceleration. The two exponential terms have been used to correct both near-site and whole-path attenuation (see Cocco and Rovelli, 1988a); the term $R$ corrects the amplitude decrement due to the geometrical spreading. The spectra $A(f)$ have been computed by Fourier transforming the two horizontal components of the acceleration time histories, using a time window which begins before the shear-wave arrival and includes the significant part of the time history. The data were windowed by using a both-end 10% cosine taper.

The estimates of the corner frequency using Equations (2) and (3) depend on the attenuation parameters $\kappa$ and $Q_o$. For $Q_o$ (seismic-$Q$ at a reference frequency of 1 Hz, assuming $Q = Q_{of}$), we used the value 100, based on the results found for Friuli in other studies (Rovelli, 1982; Rovelli et al., 1988b). The other attenuation parameter $\kappa$ is measured from the slope of the straight line fitting the high-frequency decay of the acceleration spectrum in a log-linear scale (for details, see Anderson and Hough, 1984). We have estimated $\kappa$ separately for the two horizontal component of the acceleration time histories recorded at Gemona del Friuli (see Table 1). Figure 4 shows one example of estimation of $\kappa$ for a small and a strong aftershock ($a$ and $b$, respectively). In the two cases the slope is very similar. The mean value over all data resulted in

$$< \kappa > = (0.042 \pm 0.006) \text{ sec}$$

where uncertainty represents 1 S.D.; this average value has been used in (3). Figure 4c shows the values of $\kappa$ versus earthquake size for the entire data set. We have to observe that the scatter of $\kappa$ around the mean value is similar for small and large earthquakes, even if the number of the larger events is relatively smaller. Based on this observation, we can assume that no trend exists for the high-frequency fall-off as a function of the earthquake size for this data set. This means that the effects of directivity (that can be important, mainly for the strongest events, see Boore and Joyner, 1989) are probably masked by the estimate fluctuations.

Seismic moments have been estimated directly from the spectra $A(f)$ of ground acceleration recorded at the station, assuming an omega square model $S(f; M_o, f_c)$ for source radiation

$$
A(f) = \frac{S(f; M_o, f_c) e^{-\pi \kappa f} e^{-\frac{\pi f R}{Q_o}}}{R}
$$

where $S(f; M_o, f_c)$ is defined as follows

$$
S(f; M_o, f_c) = C M_o \left( \frac{2\pi f}{f_c} \right)^2 \frac{1 + \left( \frac{f}{f_c} \right)^2}{2}
$$

$C$ is given by Equation (1). Equation (4) includes also the attenuation contribution. In (4) and (5), $M_o$ is the only unknown parameter, $f_c$ having been previously computed; the values of $\kappa$ and $Q_o$ are 0.042 and 100 sec, respectively. The inversion of the acceleration spectra in terms
Figure 4 - Example of estimation of $\kappa$ for (a) a small and (b) a strong aftershock (event 3 and 5, respectively). The slope of the high-frequency fall-off seems to be independent of the earthquake size. In (c) the values of $\kappa$ are plotted versus earthquake size. The scatter of data [the two straight lines in (c) delimitate the 1 S.D. interval around the mean value] gives an indication of the estimate fluctuation.
of \( M_0 \) was performed by applying a least-mean-square algorithm to the data in a log-log space, after correction for attenuation.

Frequency-domain estimates of \( f_c \) and \( M_0 \) were carried out for all the 24 earthquakes; their values are listed in Table 1. Figure 5 shows some specimens of acceleration spectra fitted by the theoretical model defined by (4) and (5) using the frequency domain estimates of \( f_c \) and \( M_0 \).

Table 1 summarizes time- and frequency-domain estimates of source parameters: it has to be noted that the former have not been corrected for anelastic attenuation while the latter have been corrected. The comparison between source parameters derived from these two different approaches allows to quantify to what extent anelastic dissipation is important at close distance. Moreover, it is important to observe that, while the time-domain methodology (consisting in the direct measure of the duration and the area of the displacement pulse) does not depend on the assumption of a theoretical source model, the frequency domain one needs the definition of the theoretical spectrum [an omega-square model in this work, see Equation (5)]. These topics will be discussed in the next section.

![Figure 5](image_url)  
*Figure 5 - Spectra of ground acceleration for different size earthquakes. The smooth curve represents the theoretical spectra fitting observed data, drawn on the basis of Equations (4) and (5) of the text using the frequency domain estimates of \( M_0 \) and \( f_c \) included in Table 1.*
A COMPARISON BETWEEN ESTIMATES IN TIME AND IN FREQUENCY DOMAIN

The availability of source parameters estimated in the time and in the frequency domain allows us to check both the role played by anelastic dissipation and the adequacy of the assumed spectral model [Equations (4) and (5)] including source and propagation terms. The comparison between estimates in time and frequency domain is given in Figures 6a, b and c.

**Figure 6** - Comparison between time and frequency domain estimates of source parameters. (a) Inverse of corner frequency versus source duration: the straight line represents the 1:1 scaling. (b) Seismic moments estimated in the frequency domain versus those estimated in the time domain: again the straight line represents the 1:1 scaling. The divergence from the 1:1 scaling shown by the smaller events can be interpreted in terms of the attenuation of the displacement pulse: the smooth curve represents the attenuation term $e^{-\frac{\pi F}{T}}$, that explains the deviation of data from the straight line. (c) Scaling law of source parameters estimated both in time and in frequency domain. The straight lines represent the constant stress scaling relative to different values of Brune stress drop. Frequency-domain estimates of $M_o$ and $f_c$ agree with a constant stress scaling with $\Delta\sigma = 300$ bars.
Figure 6a shows the inverse of corner frequency ($f_c^{-1}$) versus source duration $T$; the straight line represents the 1:1 scaling. The corner frequencies have been estimated in the frequency domain as discussed previously, while source durations have been estimated from the displacement time histories; in the following, we implicitly refer to $f_c$ as the frequency domain estimates, and to $T$ as the time domain ones. Figure 6a shows that for small earthquakes source durations agree with the inverse of corner frequencies; on the contrary, the same agreement is not found for the four largest aftershocks of the sequence (event 5, 6, 13 and 17 of Table 1). Two different causes may be responsible of this disagreement: the first one is an overestimation of source duration due to the greater waveform complexity observed for the largest events (see Figure 2 and 3, c and d); the second is an overestimation of corner frequency as derived from the spectrum of the near-source radiation that is mainly controlled by the size of the predominant subevent. Both these effects do not affect the estimates of source duration and corner frequency for the smaller events. Figure 6a shows also that in our data set anelastic dissipation does not significantly affect the duration of source pulse: this result is expected, as the broadening of displacement pulse $t^*$ (estimated by $\kappa$) is small compared with the source durations larger than $T \sim 0.15$ sec.

Figure 6b depicts seismic moments computed from the area of displacement pulse versus seismic moments derived from the fit of the acceleration spectra. The divergence from the 1:1 relation for small events can be explained in terms of the expected attenuation of the displacement pulse. To first order, the attenuation term $e^{-\alpha A}$ affecting the amplitude of the displacement pulse (whose duration is $T$) is sufficient to explain the deviation of time domain estimates from the 1:1 straight line (see the smooth curve in Figure 6b).

The seismic moments estimated both in time and frequency domain for the four largest events are in good agreement. For the two stronger aftershocks of September 15th, 1976 (events 13 and 17 in Table 1), teleseismic estimates of seismic moment are available (Cipar, 1980). Their values are $M_o = 8.5 \cdot 10^{24}$ dyne-cm and $M_o = 1.0 \cdot 10^{25}$ dyne-cm for the event 13 and 17, respectively, compared with $M_o = 6.0 \cdot 10^{24}$ dyne-cm and $M_o = 5.8 \cdot 10^{24}$ dyne-cm found in our study. Seismic moments obtained from strong motion data recorded at Gemona del Friuli for these two aftershocks appear to be underestimated by a factor of 1.3 and 1.7, respectively, if compared with those estimated at teleseismic distance. The better agreement between local and teleseismic estimate of seismic moment for the event 13 is probably due to the high low-frequency content in the accelerogram (Figures 2d and 3c; see also Di Bona and Boatwright, 1989). On the contrary, for the smallest earthquakes we may suppose that a reliable estimate has been obtained for both seismic moment (after correction for anelastic dissipation) and corner frequency.

These observations are summarized in Figure 6c, where the scaling law between the source parameters estimated both in time and in frequency domain has been analyzed. The straight lines represent the constant stress scaling relative to different values of Brune stress drop. This figure shows that frequency-domain estimates of $M_o$ and $f_c$ agree well with the constant stress scaling, with $\Delta \sigma = 300$ bars. The agreement includes also source parameters of the four largest events. This probably means that the underestimation of seismic moment for these earthquakes are balanced by the overestimation of corner frequency, giving estimates of Brune stress drop consistent with those obtained for the smallest events in this work, and for other large Friuli events by Cocco and Rovelli (1989). The stability of the stress drop estimates is not surprising because in our approach the observed parameter is the flat level of the ground acceleration spectrum, that is physically related to the dynamic stress drop (Boatwright, 1982 and 1988).
As a consequence of the combination of the effects shown in Figure 6a and b, time domain estimates of stress drop appear to be underestimated (roughly by a factor of two or three) if compared with frequency domain ones (see Figure 6c). This difference is probably due to the lack of attenuation compensation for the smaller events, while for the largest earthquakes it may be essentially due to the underestimation of the seismic moment, as suggested also by the comparison with teleseismic distance data (Cipar, 1980 and 1981).

The conclusion that seems to emerge from this analysis is that the anelastic dissipation plays an important role in the contamination of ground motion waveforms at close focal distance and moderate magnitudes. This will be discussed in the following sections.

**Peak Scaling Relationships**

The peak value of ground-motions is a widely used parameter both in the seismological practice (for computation of magnitudes, for example) and in the engineering community (peak acceleration has been the only routinely available measure of strong ground motion for many years). Although today other parameters can be used to characterize the seismic input in engineering practice (Fourier and response spectral values), nevertheless peak horizontal acceleration and velocity (and their scaling relationships) are still extremely important both for comparisons among data from different regions aimed at engineering purposes (Boore and Joyner, 1982; Kawashima et al., 1984; Campbell, 1985; Sabetta and Pugliese, 1987) and for more general geophysical inferences on seismogenic features and ground-motion properties (Ida, 1973; Hanks and Johnson, 1976; McGarr, 1982 and 1984; Cocco and Rovelli, 1989).

Results about scaling of peak ground motion require data from earthquakes with a wide range of magnitudes. The collection of the accelerograms from small to moderate-sized earthquakes recorded at Gemona del Friuli, although written by a conventional analog strong-motion instrument, provides a data set particularly suitable to the analysis of peak ground motion. Strong-motion data have been processed by applying a technique that provides objective and good quality displacement waveforms (see Alessandrini et al., 1989); this allowed the analysis also of peak ground displacement that is rarely available in the literature.

Figure 7 shows the scaling of peak ground acceleration, velocity and displacement versus seismic moment in the seismic moment range

$$10^{21} < M_o < 10^{25} \text{dyne cm}$$

The values plotted in Figure 7 have been computed by averaging the peak values of the two horizontal components. In the available seismic moment range, the peak values of ground motion vary significantly with earthquake size. The following relationships were obtained by fitting a regression line to the data:

$$\log a_P = -(5.98 \pm 0.69) + (0.34 \pm 0.03) \log M_o$$

$$\log v_P = -(13.23 \pm 0.76) + (0.59 \pm 0.03) \log M_o \quad (6)$$

$$\log d_P = -(20.85 \pm 0.67) + (0.88 \pm 0.03) \log M_o$$

for the peak of horizontal ground acceleration (gal), velocity (cm/sec) and displacement (cm), respectively.
Figure 7 - Scaling of peak ground acceleration, velocity and displacement versus earthquake size for the sequence of aftershocks recorded at Gemona del Friuli. Circles are relative to the seismic moment estimated in the frequency domain inverting the acceleration spectra; squares are relative to the seismic moment estimated from teleseismic recordings. The straight lines represent the regression lines obtained from the Gemona data set using frequency-domain estimates of seismic moment [Equation (6) in the text]. The dashed lines represent the Boore and Joyner's (1982) fit of peak acceleration and velocity for western North America: for sake of comparison with the Gemona data, a value \( R = 10 \) km has been used in the Boore and Joyner's regression.
In this regression, we have used the seismic moment estimated in the frequency domain as they are corrected for propagation effects. Open squares in Figure 7 represent the peak of ground motion versus teleseismic estimates of seismic moment, that are available for the four largest earthquakes. It emerges that the regression fit does not change significantly if teleseismic estimates of $M_0$ are adopted.

![Graph showing regression fit with open squares representing teleseismic estimates and full squares representing seismic moment estimated from teleseismic recordings.]

**Figure 8** - Comparison between observed peak ground motions and theoretical curves computed by means of stochastic simulations based on an omega square spectral model with constant Brune stress drop of 300 bars. Open squares are relative to the seismic moment estimated in the frequency domain inverting the acceleration spectra; full squares are relative to the seismic moment estimated from teleseismic recordings. The straight lines represent the regressions describing the behavior of peak acceleration and velocity for western North America versus seismic moment by Boore and Joyner (1982): the evidence for a factor two to three difference in peak ground motions between Friuli and California is consistent with the factor three difference in stress drop existing between the two areas.
It is noteworthy that the scaling of peak ground motion can be investigated over more than three decades of seismic moment, with similar values of the focal distance and in presence of a small impedance contrast between the sources and the receiver. The condition of similar travel path from the focus to the receiver allows us to interpret variations of peak ground motions in terms of earthquake size over a wide seismic moment range. Moreover, Figure 7 shows that peak displacements are characterized by a smaller scatter around the regression line than peak velocities and accelerations.

For sake of comparison with other regions, Figure 7 shows also the straight lines describing the behavior of peak acceleration and velocity for western North America (Boore and Joyner, 1982), computed for \( R = 10 \) km. Unlike other studies, Boore and Joyner (1982) used moment magnitude \( M \), that can be written immediately in terms of seismic moment (as defined by Hanks and Kanamori, 1979), and this makes their results directly comparable with our data. The different slope found in this work with respect to the Boore and Joyner's (1982) paper is not surprising, because of the different interval of moment magnitude where the regression has been performed: moment magnitudes range between \( 4 < M < 6 \) for our data set, against \( 5 < M < 7.7 \) for the North America data set selected by Boore and Joyner. The comparison points out the larger severity of Friuli earthquakes: Figure 7 shows that peak ground motions at Gemona for \( M_0 \approx 10^{25} \) dyne-cm assume the same values observed in California for \( M_0 > 10^{26} \) dyne-cm. Ground motions in Friuli result to be also stronger than in other Italian seismically active areas, for similar size earthquakes. This is probably due to the different tectonic environment of the Friuli region, as suggested in a recent paper (Cocco and Rovelli, 1989).

**DISCUSSIONS**

The analysis of the ground motion waveforms indicates that the behavior of source parameters in frequency domain follows a constant stress drop scaling, with a Brune stress drop of 300 bars. In this analysis, an omega-square spectral model has been assumed to describe the spectrum of seismic source radiation. The results found for source parameters in the time domain appear to be consistent with the frequency domain ones when anelastic dissipation is properly taken into account (see Figure 6).

The consistency between the peak ground motions analyzed in the previous section and the proposed scaling law needs further discussion. The stochastic simulation technique introduced by Boore (1983) offers the opportunity to analyze the behavior of peak ground motion relative to the theoretical model of the radiated spectrum. The method is based on the assumption that earthquake ground acceleration can be represented by a windowed, filtered stationary noise sequence, whose duration and spectral content have to be modelled on the basis of a theoretical model.

We generated stochastic time histories, using for the duration the inverse of corner frequency, given by

\[
  f_c = \frac{4.9 \cdot 10^6 \beta \Delta \sigma^{\frac{1}{2}}}{M_0^{\frac{3}{2}}} \tag{7}
\]

where \( \Delta \sigma \) was taken equal to 300 bars. In the simulations, we have used the spectral model (4), where the source contribution is defined by (5) jointly with (7). Simulations have been
performed for a focal distance $R = 10$ km, using $Q_0 = 100$ and $\kappa = 0.042$ sec. Figure 8 depicts the curves obtained from stochastic simulations for both peak ground acceleration and velocity. The comparison between observed data and simulations confirms the applicability, for these earthquakes, of an omega-square model with a constant stress of 300 bars. In Figure 8, the straight line fitting the western North America data (Boore and Joyner, 1982) is also shown. For both acceleration and velocity, Friuli data show a shift (roughly by a factor of two or three) toward higher peak values, in agreement with the Brune stress drop difference between the two regions (Hanks and McGuire, 1981, observed that peak ground motions in California are consistent with an average stress drop of 100 bars).

![Figure 9](image)

**Figure 9** - Scaling of ground displacement pulses for different size earthquakes: seismic moments are in the range $10^{21} < M_0 < 10^{23}$ dyne·cm.
Another interesting object of discussion is provided by the availability of good quality displacement waveforms in a wide seismic moment range. Figure 9 shows how displacement pulses scale for some of the most representative displacement time histories, in a wide amplitude range. The regular shape of the displacement pulses implies that the two parameters $d_p$ and $T$ (peak displacement and pulse duration, respectively) are in this case particularly well suited to describe the source behavior. This is possible because, for our data set, source complexity may be significant only for a few large events. In other terms, for this intermediate size data set a single pulse with amplitude $d_p$ and duration $T$ generally well represents the ground displacement radiated by the whole source process.

**Figure 10** - Scaling of the quantity $\frac{d_p}{T^2}$ versus seismic moment. The straight lines represent the constant stress scaling for different values of Brune stress drop. The apparent breakdown of source similarity is interpreted as an effect of anelastic dissipation: the smooth curve describes the joint effect of amplitude attenuation ($e^{-\phi}$) and broadening of the displacement pulse ($T + t^*$).
In particular, in the investigated seismic moment range, the quantity \( \frac{d_p}{\sigma} \) is strictly pertinent to the source scaling. In fact, recalling the relation between seismic moment and the area of the displacement pulse, we can write

\[
\frac{\Omega_v}{T^3} \propto \frac{d_p}{T^2} \propto \Delta \sigma
\]

Figure 10 contains a representation of the scaling of the quantity \( \frac{d_p}{\sigma} \) versus earthquake size. Apparently, a breakdown of similarity emerges from the data shown in this figure: the observed values of \( \frac{d_p}{\sigma} \) spread in the range from 100 to 300 bars, with a trend to decrease for decreasing earthquake size. This trend is in apparent contradiction with the finding of a constant stress drop scaling of 300 that emerges from Figure 6 and from Figure 8.

This apparent discrepancy, however, can be again explained as an effect of the anelastic dissipation on the displacement waveforms: the combination of both the amplitude decrement (roughly by a factor of \( e^{-\frac{4\pi}{5\sigma}} \) for an impulse of duration \( T \)) and the broadening of the displacement pulse (actually the observed duration is \( T + t^* \)) seems to be responsible of the trade-off exhibited by the quantity \( \frac{d_p}{\sigma} \). The smooth curve in Figure 10 represents the behavior of the following quantity

\[
\Delta \sigma \cdot \frac{T^2}{(T + t^*)^2} \cdot e^{-\frac{4\pi}{5\sigma}} = \text{const} \cdot \frac{d_p}{(T + t^*)^2} \cdot e^{-\frac{4\pi}{5\sigma}}
\]

drawn as a function of seismic moment for a constant Brune stress drop of 300 bars. This curve correctly fits the observed data: the result enhances that path attenuation has to be properly taken into account and dissipative effects cannot be neglected, even at close distances from the focus. The behavior of data shown in Figure 10 confirms the finding by Boore (1986), who demonstrated that high-frequency attenuation can be responsible of apparent breakdown of similarity in source parameter relationships for decreasing size earthquakes; \( M_o \) at which the breakdown occurs is a function of \( \kappa \). Figure 10 suggests that, for the Friuli earthquakes, anelastic dissipation affect the scaling of the quantity \( \frac{d_p}{\sigma} \) also for moderate-sized events, even if the source complexity may play a relevant role for these earthquakes. However, when anelastic dissipation effects are taken into account, the data lie around the constant 300 bars curve (see Figure 10), in agreement with all the other observations already discussed in the text.

Another important confirmation of the source parameter scaling comes from an analysis of the spectral ratios computed between events of the Gemona data set. The application of spectral ratios is particularly suitable when recordings of ground motion from a single station are analyzed, because in this case the propagation and the source contribution are intrinsically connected and cannot be easily separated. The advantage of using spectral ratios consists in eliminating propagation and site effects when the same source-to-observer path can be assumed for the earthquakes under analysis (Chael, 1987; Frankel and Wennerberg, 1989; Chun et al., 1989; Boore and Atkinson, 1989). For local earthquakes, variations in the hypocenter locations may reduce the potential capability of this methodology to eliminate propagation effects; however, spectral ratios remain the most powerful means to investigate the source scaling for earthquakes occurring close to one another.

Figure 11 shows some examples of spectral ratios for earthquakes recorded at Gemona del Friuli. The ratios have been computed using the spectrum of the smallest earthquake of the data
set (number 14 in Table 1) as the numerator (our reference spectrum). The three earthquakes whose data have been used in the denominator (event 1, 15 and 17 in Table 1) differ in seismic moment between each other slightly less than one order of magnitude; they have been chosen in order to cover all the available spectral amplitude range. Event 14 and 17 are the smallest and the largest one, respectively, while event 1 and 15 are intermediate. The theoretical curves drawn in Figure 11 have been computed relative to an omega-square model with a Brune stress drop of 300 bars.

The agreement between observed spectral ratios and theoretical curves further confirms the result of a high stress drop (300 bars) for Friuli earthquakes in a wide seismic moment range, when an omega-square model is adopted for the theoretical spectrum. The spectral ratio results are consistent with all the previous conclusions, suggesting that neither propagation or site effects significantly contaminated single-station estimates of source parameters for the Gemona data set.

**Figure 11** - Comparison between observed and theoretical spectral ratios for some of the earthquakes recorded at Gemona del Friuli. The spectrum of the smallest earthquake of the sequence (number 14 in Table 1) has been taken as the numerator. The three earthquakes whose data have been used in the denominator (event 1, 15 and 17 in Table 1) differ in seismic moment between each other slightly less than one order of magnitude. The theoretical curves have been drawn on the basis of an omega-square model with a Brune stress drop of 300 bars.
CONCLUDING REMARKS

An SMA-I accelerometer was installed at Gemona del Friuli from 22 May to the end of October, 1976. More than fifty aftershocks were recorded within a focal distance of 7 to 15 km; 24 of the records were triggered by P-waves and were particularly well suited for the study of the source scaling by analyzing the ground motion waveforms. After the digitization of the film traces by means of an automatic, high-resolution (∼ 400 sps) optical scanner, a simple numerical technique has been applied to obtain ground motion displacement directly from the digitized accelerograms by deconvolving the instrumental response in the time domain: the deconvolution operator has been analytically derived from the Laplace transform of the differential equation describing the recording system when the input is given by a Dirac function (see Alessandrini et al., 1989). Acceleration, velocity and displacement time histories have been analyzed in order to investigate the source parameter scaling in a wide seismic moment range. Source duration and seismic moment have been measured on the displacement time histories and compared with the values of corner frequency and seismic moment estimated in frequency domain from the acceleration spectra of the same events. Time domain estimates of source parameters were not corrected for anelastic attenuation: the comparison between time domain parameters and those computed in frequency domain following a methodology including spectral correction for anelastic attenuation demonstrated the bias that dissipation effects can produce in the interpretation of source properties, even at close focal distances (7 ≤ R ≤ 15 km).

After compensation of attenuation, the trend shown by source parameters both in time and in frequency domain reveals the validity of a constant stress drop scaling (Δσ ∼ 300 bars) in the investigated seismic moment range. Moreover, the peak values of the horizontal ground displacement, velocity and acceleration significantly vary with earthquake size. The scaling of peak velocity and acceleration versus seismic moment is consistent with the trend obtained from stochastic simulations assuming an omega-square spectral model with a constant stress drop of 300 bars. Interestingly, the availability of objective displacement waveforms obtained using the proposed processing technique allowed a detailed analysis of scaling of ground displacement pulse. In particular, the quantity \( \frac{d}{dt} \) has been interpreted in terms of Brune stress drop. After correcting the effects of the anelastic attenuation on the displacement pulses (amplitude decrement and broadening of duration), the scaling of the quantity \( \frac{\Delta S}{\Delta F} \) versus earthquake size is consistent with a constant stress drop of 300 bars.

The finding of this stress drop value for Friuli earthquakes is also supported by the agreement between theoretical and observed spectral ratios computed between the events of the Gemona data set. The use of this methodology is particularly important because it eliminates the effects of entire-path or near-site propagation; these effects can strongly affect estimates of source parameters derived from a single station.

The scaling laws obtained in this work agree with the results found in a previous study (Cocco and Rovelli, 1989) performed on a multi-station data set representative of the strongest earthquakes occurred in Friuli during the 1976 sequence. The analysis of the data set collected at Gemona allowed us to extend the investigation about seismic source to earthquakes with lower seismic moment, confirming also for the smaller earthquakes the validity of the results previously found by Cocco and Rovelli (1989).

Finally, this study points out the importance played by dissipation mechanisms in the analysis of ground motion waveforms, even at close distances.
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