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Observations of Short-Period Seismic Energy from Earthquakes and Inferences about the Seismic Source

David Davies and Anton Ziolkowski

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Summary

In this paper we consider the short-period radiation from earthquakes as recorded at high quality seismic stations. We pay particular attention to the profile of energy release as a function of time. Mantle phases such as P are often unsuitable for this purpose because of lateral heterogeneity, but core reflected phases are frequently much simpler.

By comparison with events of known character we are able to estimate duration of short-period energy radiation from earthquakes. Under certain assumptions, this gives us a maximum dimension D_s over which there is significant short-period energy generated. m_b 6 earthquakes with D_s as small as 2 km and as large as 50 km are described. It seems possible with an array to examine the energy release pattern of extended earthquakes in some detail.

Introduction

One of the most interesting questions that still remains in the problem of discriminating between explosions and earthquakes is the cause of the differential excitation of surface and body waves between the two types of events. A greater understanding of the $M_s : m_b$ discriminant would make it possible to predict whether there was a theoretical limit to the technique's capability at low magnitudes; it would also give us a model against which the occasional large event which fails the discriminant could be tested.

Many different possible reasons for the $M_s : m_b$ technique to be successful have been advanced. They fall into two basic categories: reasons related to spectral characteristics at the source (controlled by the displacement time function and the size of the event), and reasons related to the nature of relative excitation of P and Rayleigh waves by the two source types. In both categories it is necessary to have a much fuller theoretical and observational understanding of the earthquake source than is available at present.

There has been extensive discussion in the seismological community of the role of size in controlling spectral differences. More recently the nature of the 'time function'—a description of the displacement history of any particle on the fault plane—has also been seen to be important, although size of event and nature of the time function are probably not separable. The contribution of short period (0·5–2·0 cps) signals to this discussion has been limited. One of the reasons for this has been the complexity of the short-period signal. Rarely do earthquakes radiate simple signals to teleseismic distances, as S to P conversion near source, multipathing and scattering in a laterally heterogeneous Earth and signal generated noise near receivers all make for a large coda, from which it is impossible to extract what is ' true ' source radiation from what is consequent and irrelevant. By contrast the long-period P-wave signal has an apparent simplicity which lends itself to more immediate analysis. The simplicity however is probably deceptive. Some of the processes that lead to apparently multiple signals on short-period seismograms are equally applicable at long periods. Others, such as scattering, are strongly frequency dependent. Thus a long-period seismogram contains an indeterminate amount of energy which has travelled by an indirect route. Spectral analysis of short- and long-period seismograms produces an answer which is highly dependent on the source-receiver path and the nature of S–P conversion near source.

Conditions at the instant of an earthquake clearly change rapidly, so it is very desirable at least to make elementary measurements on the short period wave train in order to attempt to provide a more detailed description of an earthquake. To do this we shall suggest three approaches aimed at isolating the direct signal.

(i) By use of a large array we can emphasize that which is coherently propagating from a specific direction. Arrays are effective in suppressing local signal-generated noise. They are also useful in concentrating on one particular seismic phase.

(ii) By a network approach. A many-station (or better, many array) view of the seismic source should profit by having coda which is not common in type at all stations. For instance presumed explosions in Novaya Zemlya regularly produce complex signals at some stations, but there is no doubt if a network of, say, 20 stations is used that the event is very short in duration. The network is simply searched for the shortest duration signal.

(iii) By use of core phases. Phases such as PcP, PKP, $PKiKP$, and $PKKP$ are invaluable in source studies. The angle of emergence at the source is frequently within a few degrees of the vertical and S to P conversion is very small for waves almost normally incident on near-source layers. Further, the complicated multipathing effects of velocity heterogeneities in dipping slabs are generally small as few slabs appear to hang vertically beneath island arcs.

By a combination of these three techniques we shall attempt to estimate the dimensions of the source region that radiates short-period energy. At the outset however, we stress two points. We are only capable with this technique of measuring some dimension D_s associated with the radiation of short-period energy. Whether D_s is a good estimate of the fault length cannot be considered without a totally broad-band view of the source. However the resolution clearly declines as period is increased, so there is some advantage in restricting our remarks to the short-period band. It is common to associate the dominant contribution to high frequency signal to the sudden acceleration of individual points on the fault plane as the rupture passes. We should mention in addition however that the arresting of the moving rupture should also contribute in a major way to short-period energy radiation (the stopping phase, Savage 1966) and if the rupture reaches the surface we further expect a strong contribution in the short-period band (Burridge & Halliday 1971). Thus although we make the formal distinction between D_s and fault length (itself a somewhat difficult quantity to define), there is clear reason to believe that sufficient short-period energy is radiated throughout the total rupture as a continuous succession of starting, stopping and breakout phases to make D_s a very reasonable quantity to associate with an earthquake.

The second point that we wish to stress is that since we shall be using durations of signals recorded on a seismogram there is scope for subjective judgment on when a signal ends. This will always leave us with problems of crustal reverberations, and we shall see in the Turkish earthquake studied that an end is practically impossible to define. We believe nevertheless that our data do have some general interest and significance.

The earthquake model

We shall describe an earthquake as the nucleation of a rupture in a stressed medium and the outward propagation of that rupture on a fault plane until it is arrested. We make no assumptions about the fault shape—it is supposed that the outward propagation of rupture does not cease simultaneously everywhere on the fault plane. We shall assume that the rupture velocity is constant and that the time history of the displacement of any point on the fault is dominantly the sudden application of a ramp function which thereafter tends smoothly to a constant value. Energy is thus radiated as long as rupture is proceeding into new areas of the fault plane. Our model is kinematic; the dynamics of faulting do not concern us in this paper. Savage's (1966) time domain model of ' realistic faulting ' serves as a guide to the teleseismic signal to be expected from such a fault. In particular we see that the duration of the far-field elasto-dynamic signal is closely related to the largest distance that a rupture travels in the fault plane. Note that we are not assuming the far-field signal has any particular shape such as a boxcar (which would be associated with a fault model such as a narrow ribbon). Even if a fault were long and thin, we know far too little about the detailed character of faulting to assume that the fault movement is identical everywhere over the fault surface.

In Fig. 1 we show the geometrical considerations necessary to establish constraints on fault dimensions. OS is a line joining the point of nucleation to the point of termination in the direction θ relative to the direction of the observer. The length OS is a function of θ and the duration of the observed short-period signal is

$$\tau_s = \frac{\mathrm{OS}}{v}\left(1 - \frac{v}{\alpha}\cos\theta\right)$$

where v is the rupture velocity and α the P-wave velocity. We are capable only of measuring the maximum value of τ_s, which we call τ'_s.

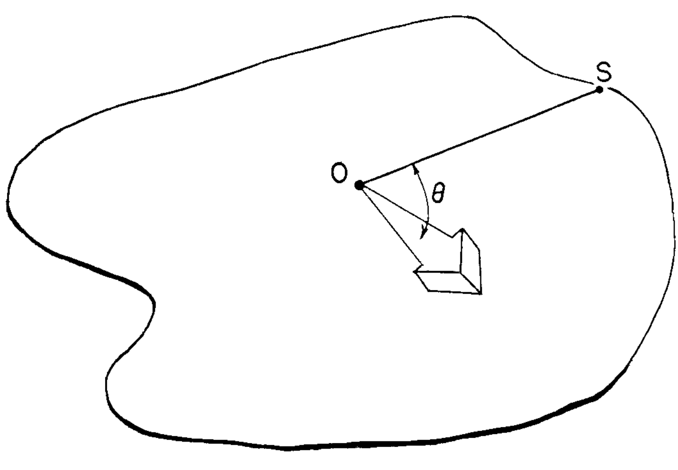

FIG. 1. Geometry of observation of a general fault. The fault plane is in the page, the observer is in the direction indicated by the arrow. Rupture is initiated at O and terminates on the closed curve of which one point is S.

We may consider as an example a strike-slip fault near the Earth's surface in which the horizontal dimension of the short-period radiator, L_s is much greater than the vertical. Since we have no *a priori* knowledge that the rupture is not bilateral, we have to assume that the dimension D_s as determined from τ_s' can only be stated to be between $L_s/2$ and L_s. For teleseismic observations θ is approximately 90°, so if we assume a rupture velocity of 3 km s^{-1},

$$3\tau'_s < L_s < 6\tau'_s.$$

For a circular shaped disc of radius r_s where rupture is initiated in the centre, the dimension estimates will clearly depend on the dip angle of the disc. For a vertical disc and teleseismic observations, assuming that v is somewhat less than β and $\alpha = \sqrt{3}\beta$ so that v/α is approximately 0·5,

$$\tau_s' = \frac{3}{2}\frac{r_s}{v}$$

since the vertically upwards travelling rupture appears to radiate the longest. A horizontal disc will yield $\tau'_s = r_s/v$, and a disc dipping at 45° (which is a common mechanism at plate boundaries) has $\tau'_s = 1{\cdot}36\,(r_s/v)$. The observed value of duration on a short-period seismogram will include instrumental response which generally lengthens the signal by about one second.

Observations of small dimension, large magnitude earthquakes

We choose first a particularly dramatic example of the use of core phases. In Fig. 2 are shown LASA beams to the *P*, *PcP* and *PKiKP* from a Nicaraguan earthquake of depth 162 km, magnitude 6·2 on 1967 Oct. 15 ($\Delta = 38°$). The mechanism for this earthquake (Isacks & Molnar 1971) has no nodal planes in the vicinity of any of these three phases on their way to LASA. The *P*-wave to LASA, however, would travel in the vicinity of the underthrust Cocos plate, although earthquakes in this region are rare below this depth so that the plate might be assumed (if mapped only by seismicity) to be well assimilated at greater depths. As it is, the signal is very complex. As we shall see later, the radiated pulse is very simple so we must assume that at least a large proportion of the complexity is generated in the source region. This may prove to be a technique for finding aseismic underthrust plates. For instance Fig. 2(a) shows a clear multi-path *P*-wave within 2 s of the start, and energy persists for at least 20 s from paths which would probably be impossible to enumerate. Beaming the array for *PcP* simplifies the signal to a degree—the small precursor disappears, confirming that it is a propagation effect, but there is still substantial energy for many seconds after the phase arrives. *PKiKP* (Fig. 2(c)) however is very much shorter, lasting no more than 3 s. We are as yet uncertain that the inner core is a truly sharp boundary (Davies & Frasier 1971) so we do not know to what extent even this 3 s of signal is indicative of structure on the inner core boundary. However, we can compare the signal with that from Milrow, an explosion of one megaton in the Aleutians whose *PKiKP* to LASA ($\Delta = 48°$) is shown in Fig. 2(d). If anything, the observed duration of the Nicaraguan event appears shorter than that of Milrow and hence τ'_s and D_s will consequently be smaller. We propose thus that $\tau'_s < 1{\cdot}0$ s and (putting $v = 5{\cdot}0$ km s^{-1} and assuming the 45° sloping disc model) that $r_s < 3{\cdot}5$ km.

We consider now a shallower event, located in the Aleutians arc. Shallow events are of greater importance in the discrimination problem because there is generally less possibility of ruling them out on depth alone. It is quite common for earthquakes in the Komandorsky Islands to exhibit long wave trains and we initially expected to

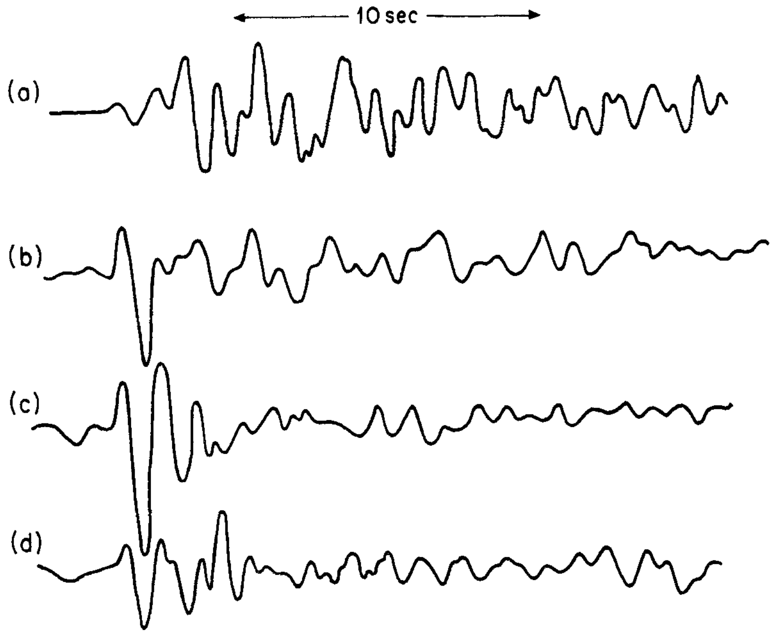

FIG. 2. Observations by LASA beams of (a) The deep Nicaraguan earthquake by *P*-waves; (b) The same by *PcP*; (c) The same by *PKiKP*; (d) Milrow, on Amchitka, by *PKiKP*.

obtain dimensions of several tens of kilometres for this event of 1969 Jan 20 with m_b 6·1, M_s 5·6, $h = 23$ km and $\Delta = 53°$ to LASA.

Fig. 3 shows this not to be the case. *PcP* is brief (*PKiKP* is similar but has a somewhat poor signal-to-noise ratio). We show in the same figure *PcP* from Longshot ($\Delta = 48°$) and it is clear, again, that this earthquake has a smaller τ_s' than the explosion with which it is being compared. We take $\tau'_s < 0{\cdot}5$ s, $v = 4\,\mathrm{km\,s^{-1}}$ and estimate r_s as less than 1·5 km.

Thus we have an earthquake of comparable dimensions, r_s, to an explosion of similar m_b but one which discriminates as an earthquake on the $M_s : m_b$ discriminant. It is thus most improbable that the enriched 20 s surface wave signal of the earthquake can be associated simply with the dimensions of the earthquake.

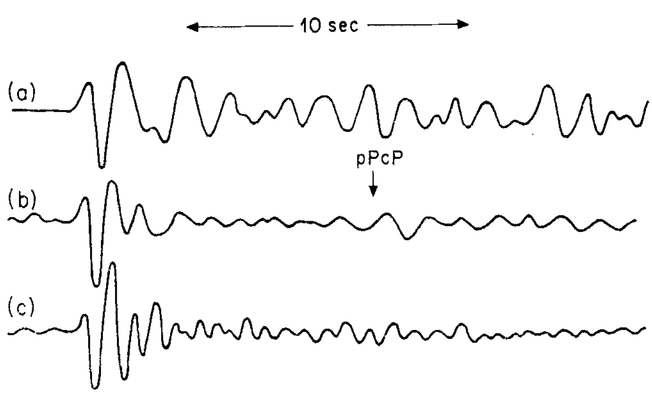

FIG. 3. Observations by LASA beams of (a) The Komandorsky earthquake by *P*-waves; (b) The same by *PcP*; (c) Longshot, on Amchitka, by *PcP*.

Observations of large dimension earthquakes

Although the two events shown thus far have had small dimensions, it should be emphasized that it is possible with arrays to look at longer durations with some degree of confidence also. One region noted for its complicated earthquakes is Turkey. It is common for P-waves from earthquakes in that region to display an emergent beginning followed several seconds later by a larger pulse. This frequently gives rise to confusion in the location process since stations operating at low gain report the second pulse, and a standard location program has to adopt a compromise between essentially two distinct data sets. Fig. 4 shows a four array view of an event in Turkey (Table 1, no. 1). This event is rich in surface waves for its quoted m_b. In addition to P at LASA it has been possible to use $PKKP$. It is clear that there is little that the teleseismic array waveforms have in common beyond an immense complexity. We take the complexity of $PKKP$ in addition to a multi-array complexity to indicate that it is the source which is drawn out in time. The Eskdalemuir record at 27° is undoubtedly confused by later phases. At the other arrays there is little to correlate. Even P and $PKKP$ at LASA and P at Yellowknife have considerable divergences although they correspond to launch angles at source differing by at most 20°. We should present one caution here before going on to discuss dimensions of this earthquake. So far we have been dealing with small-dimension earthquakes where the signal is very brief and we have been able to give an upper bound to D_s and worry little about the detailed displacement/time function for particles on the fault surface. In this case however we have at least to be aware of an alternative explanation. It is possible that the energy observed over a long time interval is radiated from a small fault which has a very complicated displacement/time function lasting for many seconds and it is this function, rather than the propagating rupture, that is being viewed teleseismically. A further alternative, of course, is that we are viewing a mixture of propagating rupture and displacement/time function.

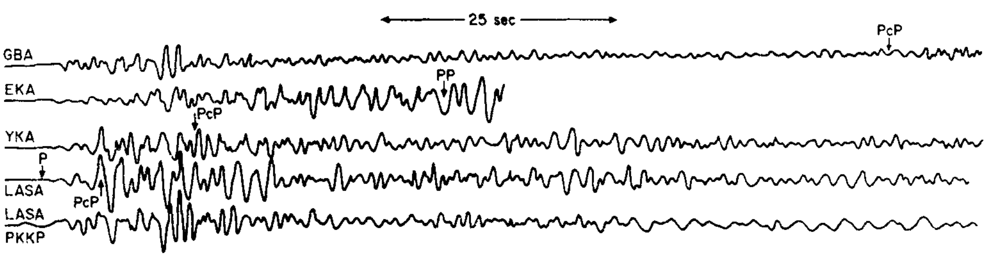

FIG. 4. Observations by arrays of the Turkish earthquake of 1968 Sept. 3. The arrays are GBA—Gauribidanur, India (△ = 48°); EKA—Eskdalemuir, Scotland (△ = 27°); YKA—Yellowknife, Canada (△ = 74°); LASA—Montana, U.S. (△ = 84°). The expected times of additional phases are marked and the ' actual ' P-wave arrival time at LASA (from Fig. 5) is also shown.

Table 1

Turkish earthquakes considered in this study (USCGS parameters except for 5 (LASA)).

	Date	Time	Location	m_b	M_s
1.	1968 Sept. 3	08 19 52	41·8° N 32·3° E	5·7	6·6
2.	1968 Sept. 3	09 13 12	41·6 N 32·3 E	4·6	—
3.	1969 Mar. 3	00 59 10	40·1 N 27·4 E	5·6	5·3
4.	1967 July 30	01 31 02	40·7 N 30·4 E	5·6	—
5.	1967 July 30	01 57 24	41·0 N 30·0 E	4·5	—

It is difficult to judge the merits of the alternatives. Physically we find it difficult to conceive of a focal process which allows for the orderly release of strain across a small surface over the duration of tens of seconds. We are particularly impressed by the rather constant energy content within the first 20 s after the brief build-up and cannot see any physical process involving such a long time constant. Nevertheless we are unable to rule this explanation out absolutely and the reader should be aware of an alternative to our discussion. The role of aftershocks is also not yet obvious.

Field observations of surface faulting would assist the discussion, but the earthquake occurred about 100 km north of the Anatolian Fault close to the Black Sea. Lander (1969) edited field reports from which there was no clear indication of surface breaks, but it seems likely that a part, at least, of the fault could have been beneath the sea. The degree of dissimilarity between observations at the different arrays *suggests*, but no more, that the source is moving. The oblique nature of this argument should give an indication of the difficulty of establishing in an uninstrumented region that the rupture truly propagated a substantial distance. Nevertheless, we shall make this assumption based on our inability to envisage a stationary long duration stress release process for a short fault.

We thus remark, based on the data shown in Fig. 4, that the duration of significant energy on all seismograms is at least 20 s, so if the rupture were uniaxial and in the Earth's crust (with low rupture velocities, say $2{\cdot}5\ \mathrm{km\,s^{-1}}$) we would have to assign a rupture length of at least 50 km to this event. Furthermore the extremely emergent beginning shown in Fig. 5 (probably only LASA amongst teleseismic stations saw the absolutely first motion) suggests that energy radiation was initially confined to a small but expanding region but rupture reached the surface, or at least alluvial layers, where a breakout occurred producing a much larger signal. Burridge & Halliday (1971) show time domain examples of this for simple models. It appears from the LASA record that once this larger phase becomes established focal processes continue to produce signal of comparable amplitude for at least 20 s and not less than half that amplitude for a further 40 s, perhaps in the same way as Wyss & Brune (1967) observed for the Alaskan earthquake. This might be taken to be propagation of the surface or suballuvial fracture. McKenzie (1972) has determined a focal mechanism for this event, which lies 100 km from the North Anatolian fault. It is not clear which portion of the event the mechanism applies to, but assuming that motion any time within the first ten seconds will be indicative of the block movement as a whole, the mechanism is dominantly overthrust with the fault plane striking NS. The nature of the faulting and its location off a major fault is reminiscent of the San Fernando earthquake of 1971.

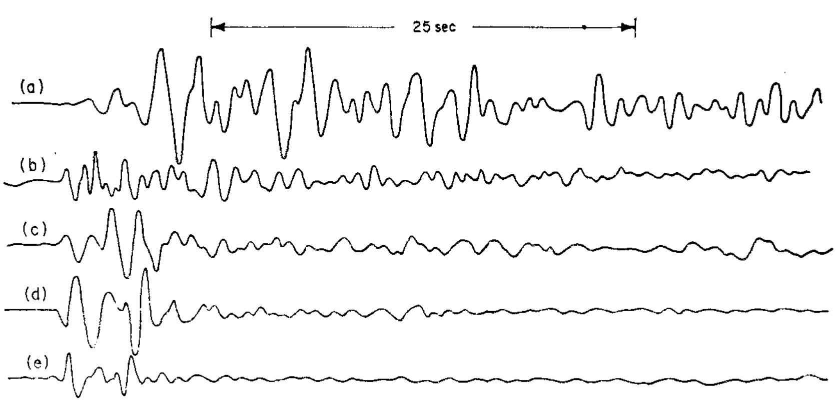

FIG. 5. LASA beams of five Turkish earthquakes, (a) 1968 Sept. 3, 0819; (b) 1968 Sept. 3, 0913; (c) 1969 Mar. 3, 0059; (d) 1967 July 30, 0131; (e) 1967 July 30, 0157.

In Fig. 5 we also show one of the many aftershocks to this event, as recorded at LASA. This one is at least one body wave magnitude smaller than the main event. It has been impossible using LASA to demonstrate any emergence such as is associated with the main shock, probably because the noise level is too high, but the duration is fully 20 s; it is likely that in this case the aftershock covered at least a large portion of the fault ruptured by the main shock. In Fig. 5 we also show recordings of two other large events from close to the event at the top of the figure, and in addition we show an aftershock (all parameters are in Table 1). It is clear from comparison of events in Fig. 5 that any lingering suspicion that a complex propagation path from Turkey to LASA *could* generate such a complex coda can be ruled out because of the complete lack of complex coda from nearby events which presumably had much smaller dimensions.

Conclusions

In order to extract source dimensions from short-period observations certain simplifying assumptions have to be made at present. We have restricted ourselves to observations in 0·5–2·0 Hz band and have made the assumption that significant energy continues to be radiated in this band for the total duration of rupture. Short-period signals are notoriously complicated by heterogeneity, but by adopting a network approach or by using reflections from the core or inner core we believe we can at least get close to an estimate for the duration of an earthquake and hence its dimensions.

We find that whilst some large earthquakes appear to be very compact, with dimensions of a few kilometres at most, Turkish earthquakes (at least) can be very complex and several tens of kilometres in size. Initiation of the rupture and subsequent 'events' are, in amplitude at least, very disparate. High quality array data should lead to an increased understanding of the history of this energy release.

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Lincoln Laboratory,
Massachusetts Institute of Technology,
Lexington, Massachusetts 02173.

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