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Determination of earthquake early warning parameters, τ_c and P_d, for southern California

Yih-Min Wu,[1] Hiroo Kanamori,[2] Richard M. Allen[3] and Egill Hauksson[2]

[1]*Department of Geosciences, National Taiwan University,* No. 1*, Sec.* 4th*, Roosevelt Rd., Taipei* 106*, Taiwan. E-mail: drymwu@ntu.edu.tw*
[2]*Seismological Laboratory, California Institute of Technology, Pasadena, CA, USA*
[3]*Seismological Laboratory, Earth & Planetary Science, UC Berkeley, CA, USA*

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SUMMARY

We explore a practical approach to earthquake early warning in southern California by determining a ground-motion period parameter τ_c and a high-pass filtered displacement amplitude parameter Pd from the initial 3 s of the P waveforms recorded at the Southern California Seismic Network stations for earthquakes with $M > 4.0$. At a given site, we estimate the magnitude of an event from τ_c and the peak ground-motion velocity (PGV) from Pd. The incoming three-component signals are recursively converted to ground acceleration, velocity and displacement. The displacements are recursively filtered with a one-way Butterworth high-pass filter with a cut-off frequency of 0.075 Hz, and a P-wave trigger is constantly monitored. When a trigger occurs, τ_c and Pd are computed. We found the relationship between τ_c and magnitude (M) for southern California, and between Pd and PGV for both southern California and Taiwan. These two relationships can be used to detect the occurrence of a major earthquake and provide onsite warning in the area around the station where onset of strong ground motion is expected within seconds after the arrival of the P wave. When the station density is high, the methods can be applied to multistation data to increase the robustness of onsite early warning and to add the regional warning approach. In an ideal situation, such warnings would be available within 10 s of the origin time of a large earthquake whose subsequent ground motion may last for tens of seconds.

Key words: earthquake early warning, magnitude, *P*-waves, seismic hazard mitigation.

1 INTRODUCTION

In the past decade, progress has been made towards implementation of earthquake early warning (EEW) in Japan, Taiwan, and Mexico (e.g. Nakamura 1988; Espinosa-Aranda *et al.* 1995; Wu *et al.* 1998; Wu & Teng 2002; Odaka *et al.* 2003; Horiuchi *et al.* 2005, 2006a). Two approaches have been adopted: (1) regional warning and (2) Onsite warning. The first approach is the traditional seismological method in which data from a seismic network are used to locate an earthquake, determine the magnitude, and estimate the ground motion in the region involved. In (2), the beginning part of the ground motion (mainly P wave) observed at a site is used to predict the ensuing ground motion (mainly S and surface waves) at the same site. In this case, it is not necessary to locate the event and estimate the magnitude, although in some systems the hypocentre and magnitude are also determined in order to estimate the strength of ground shaking in the region (Nakamura 1988; Odaka *et al.* 2003).

The regional approach is more comprehensive, but usually takes a longer time and cannot provide early warnings at distances close to the epicentre. The early warning system in Taiwan is a typical example. The Taiwan Central Weather Bureau (CWB) uses a regional warning system that requires an average of 22 s to determine earthquake parameters with magnitude uncertainties of ± 0.25. It provides a warning for areas beyond about 70 km from the epicentre. Taiwan's system has been in operation since 2002 with almost no false alarms and successfully reported many $M > 6$ events (Wu & Teng 2002; Wu *et al.* 2006a). With the advancement of new methodology and more dense seismic networks, regional systems are beginning to be able to provide early warnings to areas close to the epicentre (Tsukada & Ohtake 2002; Kamigaichi 2004; Horiuchi *et al.* 2005).

The regional approach has also been used in Japan and Mexico. The method used in Mexico is slightly different from the traditional seismological method. It is a special case of EEW system due to the relatively large distance of about 300 km between the earthquake source region (west coast of Central America) and the warning site (Mexico City). However, the warning is conceptually 'regional' (Espinosa-Aranda *et al.* 1995).

In Japan, various EEW techniques have been developed and deployed by the National Research Institute for Earth Science and Disaster Prevention (NIED) and Japan Meteorological Agency (JMA) since 2000 (Odaka *et al.* 2003; Kamigaichi 2004; Horiuchi *et al.*

2005). In particular, JMA has started sending early warning messages to potential users responsible for emergency responses. The potential users include railway systems, construction companies, and others. They are familiar with the implications of early warning messages, as well as the technical limitations of EEW. Allen & Kanamori (2003) also explored the feasibility of using data from the Southern California Seismic Network (SCSN) to issue regional early warnings.

In contrast, the onsite approach is faster, and could provide useful early warning to sites at short distances from the epicentre where an early warning is most needed. Onsite early warning can be either a single station or a dense array operation. For a single station operation, signals from P waves are used for magnitude and hypocentre determination to predict strong ground shaking. Nakamura (1984) first proposed this concept and also introduced a simple strong-motion index for onsite EEW (Nakamura 2004). However, the reliability of earthquake information is generally less than that obtained with the regional warning system. New techniques may be developed to improve the onsite warning accuracy. There is always a trade-off between warning time and the reliability of the earthquake information. Generally, an information updating procedure is necessary for any EEW system.

Onsite warning methods can be especially useful in regions where a dense seismic network is deployed. Kanamori (2005) extended the method of Nakamura (1988) and Allen & Kanamori (2003) to determine a period parameter, τ_c, from the initial 3 s of P waves. τ_c is defined as $\tau_c = 2\pi/\sqrt{r}$, where $r = [\int_0^{\tau_0} \dot{u}^2(t)dt]/[\int_0^{\tau_0} u^2(t)dt]$ ($u(t)$: ground-motion displacement; τ_0 is the duration of record used, usually 3 s, and can be computed from the incoming data sequentially. τ_c represents the size of an earthquake. Wu & Kanamori (2005a) studied the records from Taiwan systematically to explore the usefulness of the τ_c method for early warning purposes.

Using data from Taiwan, Wu & Kanamori (2005b) showed that the peak initial displacement amplitude, Pd, correlates well with the peak ground-motion velocity, PGV, at the same site. When $Pd >$ 0.5 cm, the event is most likely damaging. Wu & Kanamori (2005a,b) demonstrated that the combination of the τ_c and Pd methods can provide reliable threshold warnings within 10 s after the occurrence of a large earthquake. Wu and Zhao (2006) investigated the attenuation of Pd with the hypocentral distance R in southern California as a function of magnitude M, and obtained the following relationships:

$$M_{P_d} = 4.748 + 1.371 \times \log(P_d) + 1.883 \times \log(R) \quad \text{and}$$
$$\log(P_d) = -3.463 + 0.729 \times M - 1.374 \times \log(R). \tag{1}$$

For regional warning approach, when an earthquake location is determined by the P-wave arrival times at stations close to the epicentre, this relationship can be used to estimate the earthquake magnitude. Their result shows that for earthquakes in southern California the Pd magnitudes agree with the catalogue magnitudes with a standard deviation of 0.18 for events less than magnitude 6.5. They concluded that Pd is a robust measurement for estimating the magnitudes of earthquakes for regional early warning purposes in southern California.

In this paper, following our study for Taiwan, we have explored the use of τ_c and Pd methods for seismic early warning purposes in southern California using the data from the SCSN operated jointly by the United States Geological Survey (USGS) and the California Institute of Technology (Caltech). Fig. 1 shows the station distribution of SCSN. We will determine various regression relations between magnitude and τ_c, and Pd and PGV from the SCSN archived dataset (i.e. offline mode). The relationships determined in this paper can be used to guide the future implementation of early warning systems in real time.

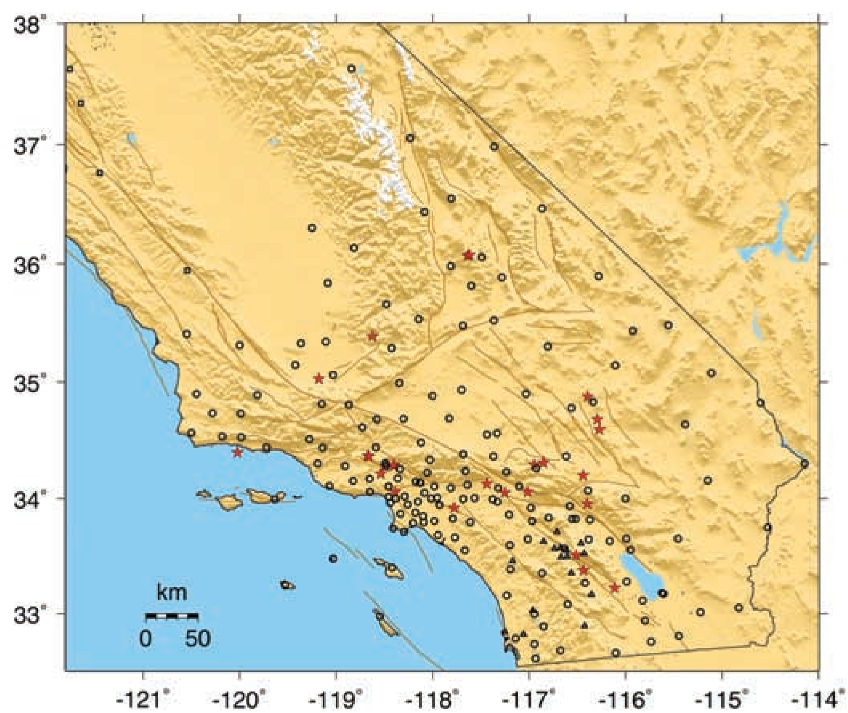

Figure 1. Map of broadband stations recorded by the Caltech/USGS Southern California Seismic Network (SCSN). SCSN stations are shown as circles, Anza Seismic Network stations as triangles, and Berkeley Digital Seismic Network (BDSN) as open squares. The epicentres of the earthquakes included in this study are shown as stars.

2 DATA AND ANALYSIS

We selected a total of 27 events in southern California listed in the SCSN catalogue for the period from 1992 to 2005. Fifteen of them are events with $M \geq 5$–7 for which our automatic *P*-wave detection algorithm picked the *P* arrival on at least three records within the epicentral distance of 100, 200 and 300 km, respectively. The remaining twelve events are with $5 > M \geq 4$ for which the *P* arrival was picked on at least 6 records within the epicentral distance of 30 km (Table 1 and Fig. 1). In the SCSN catalogue, the magnitudes are the local magnitude M_L for $M_L \leq 6$, and the moment magnitude $M_{\rm w}$ for events with $M_L \geq 6$. Here, we denote the magnitude simply by M.

Most SCSN stations have both high-gain broad-band velocity and low-gain force-balance acceleration (FBA) sensors (Hauksson *et al.* 2001). Signals are digitized at 100 or 80 samples per second with 24-bit resolution. The low-gain channels record all large earthquakes without clipping. The broad-band channels provide high signal-to-noise (S/N) ratio data for the initial 3 s of the *P* waves used in this study.

We used a total of 1161 waveforms (Table 1) in this study. Among them, 431 vertical components are used to determine τ_c and Pd. The acceleration signals are integrated to velocity and displacement. The velocity signals are differentiated to acceleration and integrated to displacement. We apply a 0.075 Hz high-pass recursive Butterworth filter to remove the low frequency drift after integration. An automatic *P* picker described by Allen (1978) is used to detect the *P* arrival from the vertical acceleration records. Since the τ_c analysis requires records with high (S/N) ratio, we set a threshold acceleration amplitude for τ_c determination. We require that the peak ground acceleration during the 3 s after the *P* arrival on the vertical component, *Pa*, be greater than 2.5 Gal. Generally, the events with small *Pa* are not important for EEW purposes. Following the analysis of Kanamori (2005), Wu & Kanamori (2005a,b), Wu *et al.* (2006b) and Wu and Zhao (2006), we use a duration of 3 s for estimation of τ_c and Pd. The longer the duration, the more reliable the magnitude estimate is, but the warning time gets shorter. The choice of 3 s is based on the numerical experiment for moment-rate functions of Sato & Hirasawa (1973) model. As shown by Fig. 3 of Kanamori (2005), if we use a duration of 3 s, τ_c increases with magnitude up to 7 beyond which it saturates. A shorter duration leads to τ_c saturation at a smaller magnitude. Although the moment rate functions of real earthquakes are different from the synthetic ones, this test suggests that a duration of 3 s yields a good measure of the magnitude of the event up to $M = 7$. Depending on the specific application, a shorter or longer duration can be used, but the trade-off mentioned above should be kept in mind.

3 RESULTS

Fig. 2 shows τ_c determined from the 431 records (Table 1) plotted as a function of magnitude, M. Although $\log\tau_c$ increases approximately linearly with magnitude, the scatter is large for events with $M <$ 5.5. The large scatter is primarily due to the low S/N ratio when the amplitude during the first 3 s is very small. However, the scatter is smaller than that in the results for Taiwan (Wu & Kanamori 2005a), probably due to the higher S/N ratio achieved by the broad-band SCSN instruments. We made τ_c determinations for the 27 events for which at least three measurements are available for each event. The three potentially damaging earthquakes with $M > 6$ all have $\tau_c > 1$ s.

Table 1. List of 27 earthquakes used in this study.

Event time		Lat.	Long.	Depth	M	Records	
		(N)	(W)	(km)		V[a]	NE[b]
1992/06/28	11:57:34.13	34.200	116.437	1.0	7.30	6	4
1994/01/17	12:30:55.39	34.213	118.537	18.4	6.70	13	4
1994/01/17	12:31:58.12	34.275	118.493	6.0	5.89	3	4
1997/04/26	10:37:30.67	34.369	118.670	16.4	5.07	15	30
1998/03/06	05:47:40.34	36.067	117.638	1.8	5.23	6	12
1998/03/07	00:36:46.84	36.076	117.618	1.7	5.04	4	8
1999/10/16	09:46:44.13	34.594	116.271	0.0	7.10	89	16
1999/10/16	09:59:35.20	34.678	116.292	10.8	5.77	9	18
1999/10/21	01:54:34.17	34.874	116.391	1.0	5.06	5	10
2000/02/21	13:49:43.13	34.050	117.252	15.4	4.37	12	24
2001/01/14	02:26:14.06	34.284	118.404	8.8	4.26	15	30
2001/01/14	02:50:53.69	34.289	118.403	8.4	4.05	15	30
2001/02/10	21:05:05.78	34.289	116.946	9.1	5.13	62	124
2001/09/09	23:59:18.04	34.059	118.388	7.9	4.24	33	60
2001/10/31	07:56:16.63	33.508	116.514	15.2	5.09	72	140
2002/01/02	12:11:28.68	33.379	116.434	12.6	4.21	11	20
2002/01/29	05:53:28.93	34.361	118.657	14.2	4.18	14	28
2002/02/22	19:32:41.75	32.319	115.322	7.0	5.70	12	22
2002/09/03	07:08:51.87	33.917	117.776	12.9	4.75	23	44
2002/09/21	21:26:16.64	33.225	116.113	14.6	4.31	7	14
2003/02/22	12:19:10.58	34.310	116.848	1.2	5.37	75	146
2004/05/09	08:57:17.30	34.395	120.022	4.4	4.40	7	12
2004/09/29	22:54:54.24	35.390	118.624	3.5	5.04	32	64
2005/01/06	14:35:27.67	34.125	117.439	4.2	4.42	17	32
2005/01/12	08:10:46.38	33.953	116.395	7.6	4.26	8	16
2005/04/16	19:18:13.55	35.027	119.178	10.3	5.15	8	16
2005/06/16	20:53:26.02	34.058	117.011	11.6	4.90	14	24

[a]Vertical component used for determination of τ_c and Pd.
[b]Both north–south and east–west components are used for determination of PGV.

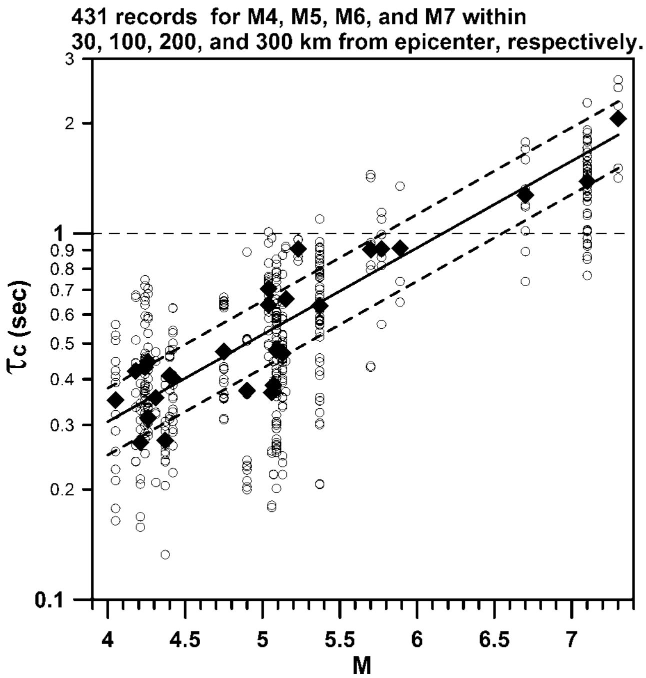

Figure 2. The τ_c measurements (open circles) from 431 records with $Pa >$ 2.5 Gal versus M. Solid diamonds show the average for the 27 events.

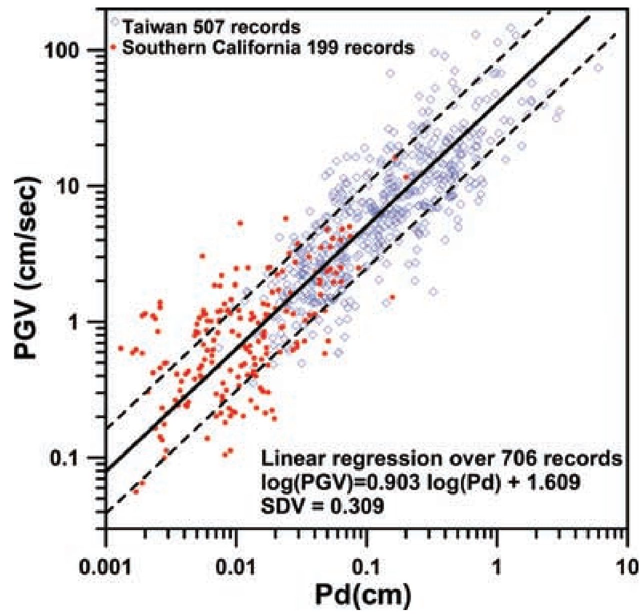

Figure 3. Relationship between peak initial displacement amplitude (Pd) measurements and peak ground velocity (PGV) for 199 records with epicentral distances less than 30 km from the epicentre in Southern California (red solid circles) and the results from Taiwan (blue diamonds). Solid line shows the least squares fit and the two dashed lines show the range of one standard deviation.

The regression with errors in both coordinates of M and τ_c results in relationships

$$\log \tau_c = 0.237M - 1.462 \pm 0.091 \quad \text{and}$$
$$M = 4.218 \log \tau_c + 6.166 \pm 0.385. \tag{2}$$

The standard deviation of the estimate of M is 0.39 for all the events. This regression is based on the average τ_c for each event with at least three measurements.

Fig. 3 shows the relationship between Pd and PGV for the 199 records with epicentral distances less than 30 km (red solid circles). PGV values increase with Pd approximately logarithmically, as observed in our previous results from Taiwan (Wu & Kanamori 2005b) shown by blue diamonds in Fig. 3. So far, the data from southern California are for relatively small ground motions. In contrast, the results from Taiwan include the data from sites with large ground motions. However, the trend is similar for Taiwan and southern California. Combining the data from Taiwan and southern California helps to extrapolate the present southern California results toward larger ground motions. For the combined data set, we obtained a regression relation

$$\log(PGV) = 0.903\ \log(Pd) + 1.609 \pm 0.309$$
$$(PGV \text{ in cm/sec and } Pd \text{ in cm}). \tag{3}$$

The southern California instrumental intensity scale for large events is defined with respect to PGV (Wald *et al.* 1999a,b). From these empirical relationships we can predict PGV or intensity from the initial 3 s of P waves for early warning purposes. In southern California, instrumental intensity (I_{MM}) is related to PGV by $I_{\mathrm{MM}} = 3.51 \log(PGV) + 2.35$ for $V \le I_{\mathrm{MM}} \le IX$ (Wald *et al.* 1999a). Thus, the standard deviation of 0.309 in log(PGV) leads to an uncertainty of 1.1 in intensity.

4 EXTENSION OF ONSITE WARNING TO REGIONAL WARNING

For onsite warning, only near-source signals are used and, in principle, the determinations of τ_c and Pd from a single station are all that are needed. When we have a very dense network, however, we can combine the data from multiple stations to increase the robustness of warning and also add the regional warning approach. We explored this possibility using 13 events selected from the list given in Table 1. These events are from 1998 to 2005 for which at least four stations within 30 km of the epicentre are available. Our algorithm automatically picks the P arrival and determines τ_c and Pd for all the records within 30 km of the epicentre. Then, we average the τ_c and Pd values determined at these stations. We use the average values to predict the average PGV for the area using eq. (3). Fig. 4 shows the relationship between the predicted and measured average PGV. The average ratio of predicted to observed PGV is 0.940 with a small uncertainty of 0.143 on the logarithmic scale. The uncertainty is much smaller than that in Fig. 3. The uncertainty of intensity prediction is now reduced from 1.1 to 0.5 units. Thus, dense array deployment can be a practical means to reduce the uncertainty of onsite EEW.

To add the regional warning approach, we use 4–6 earliest P arrivals (not every event has six stations within the distance range of 30 km) to locate the event. A simple half-space velocity model is used with a P-wave velocity of 5.8 km s^{-1}. This velocity is empirically obtained from traveltime curves within 30 km from the epicentre. Once the hypocentre is determined, we can estimate the magnitude from Pd at each station using (1), and take the average over all the stations to estimate the event magnitude. Fig. 5 shows the comparison between the magnitude estimated from Pd (denoted by M_{pd}) and the magnitude computed from τ_c(denoted by M_{τ_c}). For some of the events shown by red solid square symbols in Fig. 5, we do not have multiple stations to locate the hypocentre. For those events, we used the catalogue locations to compute M_{pd}. The

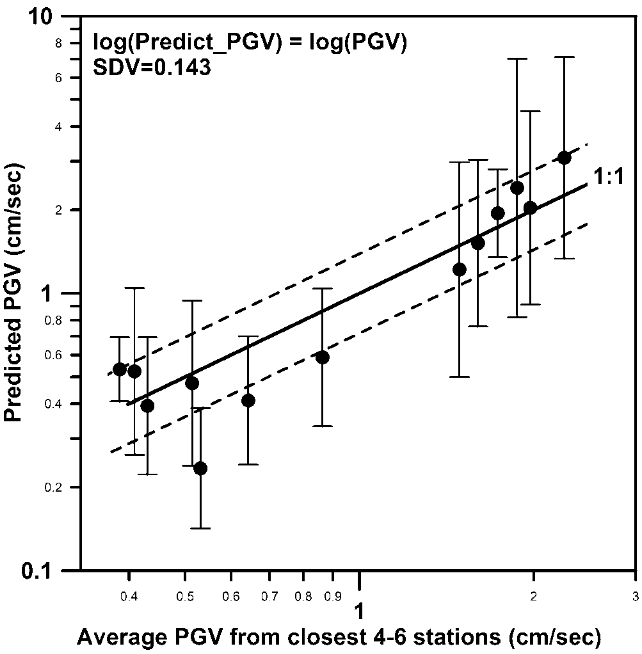

Figure 4. Relationship between predicted and observed *PGV* of 13 test events. Solid line shows 1:1 line and the two dashed lines show the range of one standard deviation.

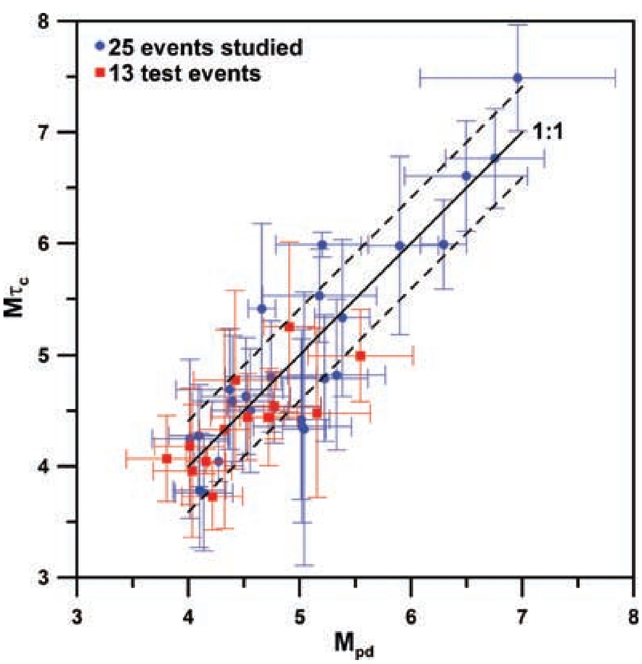

Figure 5. Magnitudes estimated from τ_c versus those estimated from *Pd*. Blue solid circles show the measurements for the 27 events. Red solid squares show the measurements for the 13 test events. Solid line shows 1:1 line and the two dashed lines show the range of one standard deviation 0.39.

agreement between M_{pd} and M_{τ_c} is good, and using *Pd*, in addition to τ_c, for magnitude estimation increases the robustness of early warning. With the hypocentre and the magnitude having been estimated, we can extend our method to regional warning, if desired. The hypocentre location error is about 6 km for the 13 events listed in Table 1. The location error of 6 km is acceptable within the point source approximation, because the source dimensions of damaging earthquakes are larger than 6 km.

5 PRACTICAL ISSUES ASSOCIATED WITH ONLINE IMPLEMENTATION

In the online mode, when τ_c, *Pa* and *Pd* are measured, no information about the event magnitude and the epicentral distance is available. As we discussed earlier, when *Pd* or *Pa* is small, the measurement of τ_c cannot be made reliably. Such an event with small *Pd* or *Pa* is not important for early warning purposes and we can ignore such events. If *Pd* or *Pa* is larger than a threshold value, the measured *Pd* and τ_c can be used to estimate the shaking intensity and *M*, respectively, although the value of *M* is not directly used for onsite early warning purposes. In principle, a single measurement of τ_c and *Pd* is enough for onsite warning, but in practice it is desirable to have more than one pair of measurements to increase the reliability and robustness of the warning. In the offline result presented above, the average of five readings provides a fairly reliable estimate. However, in the online mode, the time after the first trigger is the key parameter. If the event is located inside a dense network, then the five-station average can be obtained fairly rapidly. However, if the station distribution is sparse, it will take too long to get the five-station average to be useful for early warning. Exactly how long we should wait depends on many practical factors, and cannot be decided until actual implementation is achieved.

To establish the empirical relationships between *M* and τ_c, and *Pd* and *PGV* in real-time mode, we need to test the algorithms for a sufficiently long time. Such a testing of algorithms is presently planned within California Integrated Seismic Network (CISN).

By necessity for early warning purposes, the two parameters, τ_c and *Pd*, are determined only from the very beginning of a seismic record and considerable scatter in the τ_c versus *M* and *Pd* versus *PGV* relationships is inevitable. The errors in estimation of *M* and *PGV* may lead to false alarms and missed alarms. How to deal with false alarms and missed alarms is an important issue in EEW, but it must be discussed in relation to the specific use of early warning information, which is beyond the scope of this paper.

6 DISCUSSIONS AND CONCLUSIONS

The period parameter τ_c from SCSN is smaller than that for a similar magnitude earthquake in Taiwan. For most events with $M > 5$ in Taiwan, $\tau_c > 1$ s while for the events with $M > 6$ in southern California, $\tau_c > 1$ s. This may be due to the difference in S/N ratio, especially for smaller earthquakes. Records from Taiwan are 16-bit strong-motion data, and those from SCSN are mainly from 24-bit broad-band data, especially for small earthquakes. The use of strong-motion data involves double integration which occasionally results in long-period drift and larger τ_c. However, the difference in τ_c is expected to decrease as the event size increases. For practical early warning purposes, we are concerned with the events with $M > 6$.

Considerable uncertainties in magnitude determination or intensity prediction using the initial *P* waves are inevitable. If the seismic network is dense we can significantly reduce the uncertainties by taking the average of the measurements from several stations. In our test, the uncertainties of magnitude determination and intensity can be reduced to 0.3 and 0.5, respectively, if more than four stations are available within 30 km of the epicentre.

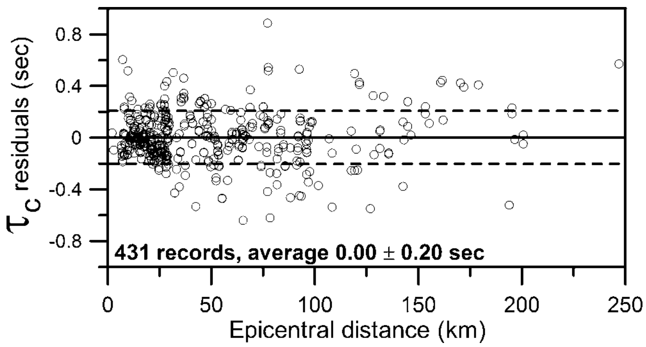

Figure 6. τ_c residuals (deviation of τ_c from the average of each event) versus hypocentral distance for the 431 records with $Pa > 2.5$ Gal.

In our model, τ_c is a source parameter and should not depend on the distance. To test this, we plot in Fig. 6 the deviation of τ_c from the average τ_c for each event as a function of the hypocentral distance. In total, 431 records with $Pa > 2.5$ gal from the nine events in Table 1 are used. At the distance less than 150 km, τ_c does not depend on distance, but τ_c tends to be positive at distances larger than 150 km. It is possible thatτ_c increases with distance as the waves are more attenuated. In practical applications, we measure τ_c at short distances, mostly less than 100 km, and the distance dependence can be ignored.

From the results of our off-line experiment, we estimate that the earthquake parameters can be typically determined at about 10 s after the origin time. The 10 s is the sum of the *P*-wave traveltime to the last reporting station, the 3 s over which we measure τ_c and *Pd*, and the potential delay time, about 3 s, during signal transmission of the current SCSN. Fig. 7 shows the *P*- and *S*-wave traveltimes for an event at a reference focal depth of 5 km. Since *S* waves propagate to sites about 30 km from the epicentre in 10 s, if the method is implemented online, we can provide early warnings to the sites farther than 30 km from the epicentre. The warning information could include earthquake location, magnitude and estimated intensity. Because the SCSN system was not specifically designed for early warning purposes, considerable delays occur during data processing. If the delays in the system including those during signal transmission are shortened in the future, early warning information may be available at sites with a distance shorter than 30 km. For deeper events such as the 1994 Northridge earthquake that had a hypocentral depth of 16 km, the minimum distance where early warning is available can be even shorter.

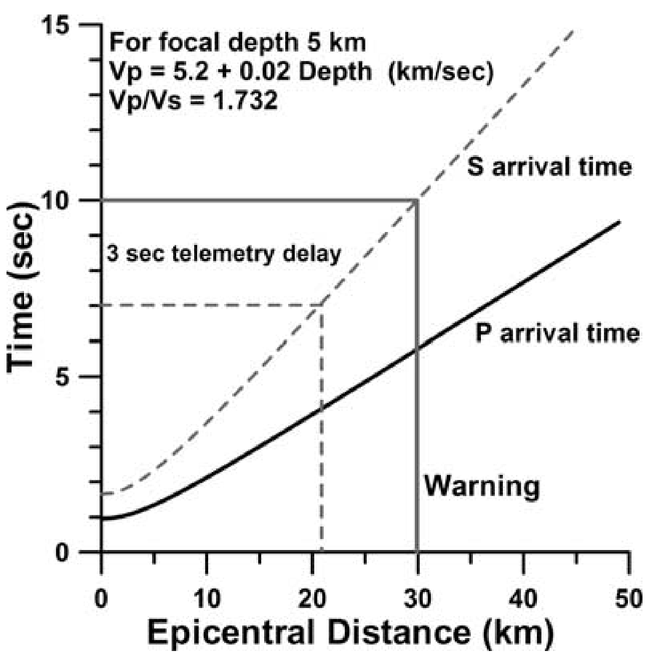

Figure 7. Reporting time and regional warning distance achievable by the combined τ_c and *Pd* methods.

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