**M_{wpd}: A Duration-Amplitude Procedure for Rapid Determination of Earthquake Magnitude and Tsunamigenic Potential from P Waveforms**

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**Summary**

We present a duration-amplitude procedure for rapid determination of an earthquake moment magnitude, $M_{wpd}$, from P-wave recordings at teleseismic distances. The $M_{wpd}$ magnitude can be obtained within 20 minutes or less after the event origin time since the required data is available in near-real time. The procedure determines apparent source durations, $T_0$, from high-frequency, P-wave records, and estimates moments through integration of broadband displacement waveforms over the interval $t_P$ to $t_P + T_0$, where $t_P$ is the $P$ arrival time. We apply the duration-amplitude methodology to a number of recent, large earthquakes (Global Centroid-Moment Tensor magnitude, $M_{w}^{CMT}$, 6.6 to 9.3) with diverse source types. The results show that a scaling of the moment estimates for interplate thrust and possibly tsunami earthquakes is necessary to best match $M_{w}^{CMT}$. With this scaling, $M_{wpd}$ matches $M_{w}^{CMT}$ typically within ±0.2 magnitude units, with a standard deviation of $\sigma$=0.10, outperforming other approaches to rapid magnitude determination. In addition, $M_{wpd}$ does not exhibit saturation for the largest events, or, equivalently, $\Delta M$=$M_{wpd}$-$M_{w}^{CMT}$ does not become more negative with increasing $M_{w}^{CMT}$. The explicit use of the source duration for integration of displacement seismograms, the moment scaling, and other characteristics of the duration-amplitude methodology make it an extension of the widely used, $M_{w}$ rapid-magnitude procedure. The obtained durations and duration-amplitude moments allow rapid estimation of the energy-to-moment ratio $\Theta$ used for identification of tsunami earthquakes. The need for a moment scaling for interplate thrust and possibly tsunami earthquakes may have important implications for the source physics of these events.

**Key words:** earthquakes, Richter magnitude, seismic moment, seismograms, tsunami, earthquake-source mechanism.

**Introduction**

Effective tsunami warning and emergency response for large earthquakes requires accurate knowledge of the event size within 30 minutes or less after the event origin time (OT). The 26 December 2004, M9 Sumatra-Andaman earthquake caused a tsunami that devastated coasts around the Eastern Indian Ocean within 3 hours; the 17 July 2006, M7.7 Java earthquake
caused an unexpectedly large and destructive tsunami. For both events the magnitudes available within the first hour after the event origin time severely underestimated the event size (Kerr, 2005; PTWC, 2006ab).

Currently, the earliest, accurate estimates of the size of major and great earthquakes come from moment tensor determinations, including the authoritative, Global Centroid-Moment Tensor (CMT) determination and corresponding moment-magnitude, $M_{w}^{CMT}$ (Dziewonski et al., 1981; Ekström, 1994), and related procedures (e.g., Kawakatsu, 1995). These estimates are based on long-period, seismic S and surface-wave waveform recordings, typically not available until an hour or more after OT.

The mantle magnitude, $M_{m}$ (Okal and Talandier, 1989; Newman and Okal, 1998; Weinstein and Okal, 2005) is also based on surface waves. The spectral amplitude of mantle Rayleigh waves at variable periods (between 50 and 300 sec for large events), combined with approximate corrections for geometrical spreading and Rayleigh wave excitation at the source, gives the $M_{m}$ estimate and a corresponding moment. $M_{m}$ is potentially available within minutes after the first Rayleigh wave passage (i.e. about 20 min after OT at 30° great-circle distance (GCD), and about 50 min after OT at 90° GCD), but for very large events the analysis of waves at increased periods (450 sec or more) may be required (Weinstein and Okal, 2005; UNESCO, 2005) leading to an increased delay after OT for obtaining the $M_{m}$ estimate.

Seismic P-waves are the first signals to arrive at seismic recording stations. At teleseismic distances (30-90° GCD) the arrival times of the initial P-wave are used routinely to locate the earthquake hypocentre within about 10 to 15 minutes after OT. The initial P-waves and following P-wave train also contain comprehensive information about the event size and source character. Boatwright and Choy (1986) show that the total radiated seismic energy can be estimated from the P-waves alone.

There are a number of procedures for rapid analysis of large earthquakes using seismic P-waves currently in use at earthquake and tsunami monitoring centers. Because these procedures use only the P-wave portion of a seismogram, event size estimates are potentially available only a few minutes after the P wave has been recorded at teleseismic distances, i.e. in as little as 10-15 min after OT at 30° GCD, and about 20 min after OT at 90° GCD.

The U.S. Geological Survey National Earthquake Information Center (NEIC) Fast Moment Tensor procedure (Sipkin, 1994; http://earthquake.usgs.gov) produces an estimate of the seismic moment tensor and moment magnitude, $M_{w}^{NEIC}$, for earthquakes of magnitude of 5.5 or greater within the order of 30 min after OT through automated processing and inversion of P-wave waveforms.

The widely used, $M_{sp}$ moment-magnitude algorithm (Tsuboi et al., 1995; Tsuboi et al.,1999; Tsuboi, 2000) considers very-broadband, P-wave displacement seismograms as approximate far-field, source-time functions. These displacement seismograms are integrated and corrected approximately for geometrical spreading and an average radiation pattern to obtain scalar moments at each station. Application of the standard moment magnitude formula, averaging over stations and optionally applying a magnitude dependent correction (Whitmore et al., 2002) gives a moment magnitude, $M_{sp}$, for an event.

$M_{w}^{NEIC}$ and $M_{sp}$ match closely $M_{w}^{CMT}$ up to $M_{w}^{CMT} \approx 7.5$, but at greater magnitudes they tend to increasingly underestimate $M_{w}^{CMT}$ (Figure 1, Table 1). To resolve this magnitude saturation problem while providing accurate and rapid magnitude estimates for large earthquakes, a number of authors have proposed new methodologies for magnitude determination based on
P-wave signals.

Menke and Levin (2005) propose that the ratio of long-period, P-wave displacement amplitudes between an target event and a nearby reference event of known size can rapidly provide the magnitude of the target event. Lockwood and Kanamori (2006) show that wavelet analysis of P-waves distinguishes a significantly greater amplitude of the long-period, W phase for the 26 December 2004, M9 Sumatra-Andaman relative to the W phase of the 28 March 2005, M8.6 Northern Sumatra earthquakes. They propose that this procedure can be used for rapid identification of the largest, great earthquakes and their high tsunamigenic potential.

Bormann and Wylegalla (2005) and Bormann et al. (2006) calculate a cumulative $m_B$ magnitude, $m_{Bc}$, by summing up the peak velocity amplitudes for all pulses (signal between two consecutive zero crossings) in the P waveform. Hara (2007) combines measures of the high-frequency duration and maximum displacement amplitude of P-waveforms for a set of large, shallow earthquakes to determine an empirical relation for moment magnitude.

Lomax (2005) shows for very large earthquakes that the location of the end of rupture, and thus an estimate of the event size, can be rapidly determined from measures of the P-wave duration on high-frequency records. Lomax and Michelini (2005) note that the ratio of the high-frequency, P-wave durations from the 2004, M9 Sumatra-Andaman and the 2005, M8.6 Northern Sumatra earthquakes match the ratio of the CMT moment values for the two events, and suggest that the high-frequency, P-wave duration could be used for rapid magnitude estimation for individual events. Lomax et al. (2007) use teleseismic (GCD $\geq 30^\circ$), P-wave signals to estimate radiated seismic energy, $E$, and source duration, $T_0$, and show that an energy-duration moment relation, $M_0^{ED} \propto E^{1/2} T_0^{3/2}$, based on an expression for $E$ from Vassiliou and Kanamori (1982), gives a moment magnitude, $M_{ED}$, that matches closely $M_w^{CMT}$ for a set of recent, large earthquakes.

These new methodologies for rapid magnitude determination based on P-wave signals all produce useful magnitude estimates, $M_{est}$, for very large earthquakes. Most of these methodologies, however, show significant differences with $M_w^{CMT}$ (i.e., $|M_{est} - M_w^{CMT}| \geq 0.3$) for many events, including some of the most important and destructive interplate thrust events and tsunami earthquakes (tsunami earthquakes are characterized by unusually large tsunamis and a deficiency in moment release at high frequencies, e.g., Kanamori, 1972; Polet and Kanamori, 2000; Satake, 2002). Most of these methodologies also give $\Delta M = M_{est} - M_w^{CMT}$ values which becomes more negative with increasing $M_w^{CMT}$; this effect is equivalent to the magnitude saturation of $M_w^{NEIC}$ and $M_{sp}$.

To further investigate and resolve these problems, we introduce here a rapid and robust, duration-amplitude procedure to obtain an earthquake moment and a moment magnitude, $M_{spa}$, from P-wave recordings at teleseismic distances. This procedure first determines apparent source durations, $T_0$, from high-frequency, P-wave records, and then estimates moments through integration of broadband displacement records over the $t_P$ to $t_P + T_0$ interval, where $t_P$ is the P arrival time. This methodology can be viewed as an extension of the $M_{sp}$ moment-magnitude algorithm. We apply the duration-amplitude procedure to a number of recent, large earthquakes with diverse source types.

**Improved estimation of seismic moment from P waveforms**

Given the far-field, P-displacement waveform, $u(t)$, for an earthquake source of duration, $T_0$,
then a theoretical expression for the scalar, seismic moment, $M_0$, is,

$$M_0 = C_M \int_{t_p}^{t_p+T_0} u(t) \, dt,$$

(1)

where $t_p$ is the $P$ arrival time, $u(t)$ is corrected for geometrical spreading and attenuation, and $C_M$ is a constant that depends on the density and wave speed at the source and station, a double-couple radiation pattern and other factors (e.g., Aki and Richards, 1980; Boatwright and Choy, 1986; Tsuboi et al., 1995; Newman and Okal, 1998; Kanamori and Rivera, 2004; see Appendix A for details). Equation (1) suggests that the scalar moment, $M_0$, of an earthquake can be determined from $P$ wave displacement seismograms. Application of the standard moment-magnitude formula to the obtained $M_0$,

$$M_w = (\log_{10} M_0 - 9.1) / 1.5,$$

(2)

(Hanks and Kanamori, 1979) gives a $P$-wave estimate of the moment magnitude, $M_w$, for an event.

Equation (1) cannot be used directly to obtain accurate moment estimates for a number of reasons, including the presence of surface reflected and other secondary phases, and the difficulty of estimating $T_0$. The $M_{wp}$ magnitude procedure addresses some of these problems by estimating the scalar moment from either the first peak or the first peak-to-peak amplitudes on $P$-displacement seismograms integrated using Equation (1), though the integral is performed without explicit knowledge or use of $T_0$ (e.g., Tsuboi et al., 1999).

To make further use of Equation (1) to obtain more accurate, rapid moment-magnitude estimates, we begin by examining moments, $\hat{M}_0$, and magnitudes, $M_{wpd}$, determined through application to teleseismic $P$-displacement seismograms of a modified form of Equation (1),

$$\hat{M}_0 = k C_M \text{Max} \left[ \int_{t_p}^{t_p+T_0} u(t) \, dt, \int_{t_p}^{t_p+T_0} |u(t)| \, dt \right].$$

(3)

The modifications in Equation (3) includes the following: 1) The integral in Equation (1) is taken separately over the positive, $u^+(t)$, and the absolute value of negative, $|u^-(t)|$, displacement amplitudes to help separate the direct $P$ waves from surface reflection phases and other phases with opposite polarity; the maximum of these two integrals is used to calculate the moment estimate. 2) A regression constant, $k$, is included to compensate for unknown errors and biases in the terms of $C_M$ and in the correction of $u(t)$ for attenuation and geometrical spreading. In addition, the source duration, $T_0$, is estimated through measures on high-frequency, $P$-wave seismograms (Lomax, 2005; Lomax et al., 2007) and explicitly used to define the upper limit of integration. Application of the standard moment-magnitude formula, Equation (2), using $\hat{M}_0$ and averaging over stations using robust statistics gives a $P$-wave moment magnitude, $M_{wpd}$, for an event.

Further details on this procedure are given in Appendices A, B and C; the processing steps are illustrated in Figure 3. We note here that the amplitude correction of the displacement waveforms for attenuation and geometrical spreading and the calculation of $C_M$ make use of the PREM model (Dziewonski and Anderson, 1981) without a crust (hereinafter referred to as PREM_NC), since most large events occur in oceanic regions. For shallow continental events, the effect of the crust on $C_M$ is introduced as a magnitude correction using the PREM properties for the lower crust. Also, the radiation pattern factor in $C_M$ for strike-slip events,
which differs from that for all other event types, is determined empirically. Table 1 indicates
the classification of each event according to source type and oceanic versus continental
setting, mainly based on event information from the NEIC (http://earthquake.usgs.gov) and
the Global CMT Catalog (http://www.globalcmt.org). This classification takes into account
the epicenter, depth and moment-tensor mechanism in relation to the background seismicity
and the surrounding tectonic plates and plate boundaries; in a few cases additional
information from the NEIC tectonic summary is used.

Figure 4 shows a comparison of the obtained magnitudes, $M_{\text{wpd}}$, with $M_{\text{w,CMT}}$ for 79, recent,
large earthquakes ($M_{\text{w,CMT}}$ 6.6 to 9.3; Figure 2 and Table 1). This comparison shows that $M_{\text{wpd}}$
matches closely $M_{\text{w,CMT}}$ up to $M_{\text{w,CMT}} \sim 7.5$, but with increasing magnitude $M_{\text{wpd}}$ tends to
increasingly underestimate $M_{\text{w,CMT}}$. This is a similar result as obtained for $M_{\text{wp}}$ (Figure 1),
though $M_{\text{wp}}$ gives an even larger underestimate than $M_{\text{wpd}}$ of $M_{\text{w,CMT}}$ above $M_{\text{w,CMT}} \sim 7.5$,
primarily because $M_{\text{wp}}$ only considers the first part of the $P$ wave train while $M_{\text{wpd}}$ is based on
the full interval of duration $T_{\text{p}}$ after the $P$ arrival. The NEIC Fast Moment Tensor magnitude,$M_{\text{w,NEIC}}$, (Sipkin, 1994; http://earthquake.usgs.gov), based on waveform inversion, also shows
an increasing underestimate of $M_{\text{w,CMT}}$ above $M_{\text{w,CMT}} \sim 7.5$ (Figure 1).

Closer examination of Figure 4 shows that the trend of increasing underestimate of $M_{\text{w,CMT}}$ by
$M_{\text{wpd}}$ (i.e. $\Delta M = M_{\text{wpd}} - M_{\text{w,CMT}}$ becomes more negative) with increasing $M_{\text{w,CMT}}$ occurs mainly for
interplate thrust earthquakes (type I in Table 1). $M_{\text{wpd}}$ agrees well with $M_{\text{w,CMT}}$ for most events
of other types, agreeing over a wide range of magnitudes for strike-slip (types S and So),
intraplate (type P), intermediate depth (downdip, type W) and deep earthquakes (type D), and
over the limited range of available magnitudes for reverse-faulting (type R and Ro) and
normal-faulting (type N and No) crustal earthquakes. The lack of large events of tsunami
(type T), however, prevents excluding that this types follow a trend similar to that of interplate
thrust earthquakes.

Thus we find for larger ($M_{\text{w,CMT}} > \sim 7.5$) interplate thrust events that the moments determined
from the $P$-wave train through application of Equation (3), and apparently also through
waveform inversion (e.g., $M_{\text{w,NEIC}}$, Figure 1a), underestimate the corresponding CMT
moments, derived from inversion of long period S and surfaces wave.

### Moment scaling for interplate thrust and tsunami earthquakes

The variation of $\Delta M = M_{\text{wpd}} - M_{\text{w,CMT}}$ differences for interplate thrust earthquakes as a function of
$M_{\text{w,CMT}}$ (Figure 4b) and a similar variation as a function of $M_{\text{wpd}}$ suggest that more accurate
moment estimates for these events, $M'_0$, can be obtained by scaling $\hat{M}_0$ with a factor
composed of $\hat{M}_0$ raised to some power, i.e.,

$$M'_0 = \hat{M}_0 \left( \frac{M'_0}{M_{0,\text{cutoff}}} \right)^R,$$  \hspace{1cm} (4)

where $\hat{M}_0$ is given by Equation (3) and $M_{0,\text{cutoff}}$ is a constant cutoff moment below which the
scaling is not applied. We also apply the moment scaling, Equation (4), to tsunami
earthquakes, since these events fall within the trend of $\Delta M$ differences for interplate thrust
earthquakes and because it is difficult to distinguish these two types of events in near real-
time analysis. Application of the standard moment-magnitude formula, Equation (2), and
averaging over stations gives the corresponding $P$-wave moment magnitude, $M_{\text{wpd}}$. (see
Appendix B for further details)
Application of Equation (4) to the interplate thrust and tsunami events from the set of studied earthquakes using values of $R$ in the range $0 \leq R \leq 1.3$ gives $R \approx 0.4$ and $M_0^{\text{cutoff}} \approx 7.5 \times 10^{19}$ N-m (equivalent to $M_0 \approx 7.2$) for the best match of $M_{wpd}$ to $M_w^{CMT}$. (The optimal value of $M_0^{\text{cutoff}}$ and $R$ are sensitive to the algorithms used to estimate $T_0$ and moment, see Appendix B). Thus we arrive at a preferred, duration-amplitude expression for moment estimation,

$$M_0^{pd} = \tilde{M}_0 \left( \frac{M_0}{M_0^{\text{cutoff}}} \right)^{0.4}, \quad (5a)$$

for interplate thrust and tsunami events with $\tilde{M}_0 \geq M_0^{\text{cutoff}}$, and

$$M_0^{pd} = \tilde{M}_0, \quad (5b)$$

otherwise, where $\tilde{M}_0$ is given by Equation (3) with $C_M = 1.62 \times 10^{19}$ and $k = 1.1$. $M_{wpd}$ magnitudes determined using Equations (5a), (5b) and (2) for the studied earthquakes match closely $M_w^{CMT}$ for all event types at all magnitudes (Figure 5).

**Estimation of energy-to-moment ratio $\Theta$**

The energy-to-moment ratio parameter, $\Theta$, (e.g., Newman, and Okal, 1998; Weinstein and Okal, 2005) for identification of tsunami earthquakes is defined as,

$$\Theta = \log_{10} \frac{E}{M_0}, \quad (6)$$

where $E$ is the radiated seismic energy and $M_0$ the moment. For most earthquakes, this parameter is expected to have a value of $\Theta \approx -4.9$, but $\Theta$ values around -6.0 or less are found for tsunami earthquakes (Weinstein and Okal, 2005). Thus anomalously low values of a rapid estimate of $\Theta$, combined with knowledge of an earthquake’s location, size, tectonic setting and likely source type, can be an important indicator of a potential tsunami earthquake.

From duration-amplitude estimates of moment, $M_0^{pd}$, and duration, $T_0$, we can obtain $\Theta$ through application of the energy-duration relation of Lomax et al. (2007),

$$M_0^{ED} = c \frac{E^{1/2} T_0^{3/2}}{M_0}, \quad (7)$$

where $c \approx 1.55 \times 10^{10}$ for average crust - upper mantle material properties. Substituting $M_0^{pd}$ for $M_0^{ED}$ in Equation 7, solving for $E$ and substituting into Equation 6, we get,

$$\Theta = \log_{10} \left( c^{2} \frac{M_0^{pd}}{T_0^3} \right). \quad (8)$$

$\Theta$ values determined using Equation (8) for the studied earthquakes are listed in Table 1 and plotted in Figure 6 as function of $M_w^{CMT}$.

**Discussion and Conclusions**

We have introduced a duration-amplitude procedure to obtain rapidly an earthquake moment, $M_0^{pd}$, and moment magnitude, $M_{wpd}$, from $P$-wave recordings at teleseismic distances. Because the required recordings are available in near-real time at earthquake and tsunami monitoring centers, $M_{wpd}$ can be available within about 20 minutes after OT. For major and great earthquakes ($M_w^{CMT} \geq 7.0$), $M_{wpd}$ (with moment scaling for interplate thrust and tsunami events) matches $M_w^{CMT}$ typically within $\pm 0.2$ magnitude units, with a standard deviation of
only $\sigma=0.10$ (Figure 5, Table 1). In addition, $M_{wpd}$ does not exhibit saturation for the largest events, or, equivalently, $\Delta M = M_{wpd} - M_w^{CMT}$ does not become more negative with increasing $M_w^{CMT}$. Thus $M_{wpd}$ outperforms other procedures for rapid moment magnitude determination. The results of other procedures, using different and smaller sets of events than used here, are: $M_{ED}$ (Lomax et al., 2007) matches $M_w^{CMT}$ typically within $\pm 0.3$ magnitude units, with $\sigma=0.16$, and $\Delta M$ for $M_{ED}$ does not become more negative with increasing $M_w^{CMT}$; $mBc$ (Bormann et al., 2006) matches $M_w^{CMT}$ typically within $\pm 0.5$ magnitude units, with $\sigma=0.26$, and $\Delta M$ for $mBc$ becomes more negative with increasing $M_w^{CMT}$; the rapid magnitude estimates of Hara (2007) shows a match with $M_w^{CMT}$ typically within $\pm 0.3$ magnitude units, with $\sigma=0.18$, and $\Delta M$ for this magnitude is stable or possibly becomes more negative with increasing $M_w^{CMT}$, our corrected $M_{sp}$ results (Figure 1c; Table 1) match $M_w^{CMT}$ typically within $\pm 0.5$ magnitude units, with $\sigma=0.25$, and $\Delta M$ for $M_{sp}$ becomes rapidly more negative with increasing $M_w^{CMT}$.

The improved agreement between $M_{wpd}$ and $M_w^{CMT}$ relative to other rapid procedures, including $M_{wp}$, can be attributed primarily to the use in Equation (3) of the full $t_p$ to $t_p+t_0$ interval for integration with testing of integrals over positive and negative values of displacement, and to the application of the moment scaling, Equation (5a), for interplate thrust and tsunami earthquakes. This agreement is also dependent on the use of a robust procedure for estimating $T_0$ from high-frequency seismograms, and of additional corrections for certain events types (see Appendices A and B for details). The $M_{wpd}$ results indicate that testing of the integral in Equation (3) over positive and negative values of displacement separates adequately the direct $P$ waves from surface reflection phases and other secondary phases, even when the rupture duration, $T_0$, is large. This testing is analogous to the selection in the $M_{sp}$ magnitude procedure of the larger of the first peak or the first peak-to-peak amplitude of the integral Equation (1). The moment scaling used here is likely related to the magnitude dependent correction to $M_{sp}$ proposed by Whitmore et al. (2002) and to the values of the coefficients in the regression of Hara (2007), in both cases applied to all earthquakes. In contrast, we find here that moment scaling is only needed for interplate thrust earthquakes, and possibly for tsunami earthquakes. The characteristics of the duration-amplitude procedure noted above show that it is an extension of the $M_{sp}$ moment-magnitude algorithm, recalling also that both procedures are ultimately based on Equation (1).

**Energy-to-moment ratio $\Theta$**

We have shown that the duration-amplitude estimates of moment, $M_{spd}$, and duration, $T_0$, can be combined with the energy-duration relation of Lomax et al. (2007) to provide a rapid estimate (Equation 8) of the energy-to-moment ratio $\Theta$ (e.g., Newman, and Okal, 1998; Weinstein and Okal, 2005) used for identification of tsunami earthquakes. Duration-amplitude estimates of $\Theta$ estimated using Equation 8 are listed in Table 1 shown in Figure 6, along with two lines of constant $\Theta$: $\Theta = -4.9$, the expected value for all earthquakes, and $\Theta = -5.5$, below which indicates a possible tsunami earthquake (e.g., Weinstein and Okal, 2005). The duration-amplitude estimates of $\Theta$ are $\Theta \leq -5.9$ for all tsunami earthquakes, facilitating identification of these events. Some interplate thrust events and many strike-slip events have low $\Theta$ values below -5.5, and most deep events have high $\Theta$ values. $\Theta$ is low, $\Theta = -6.1$, for a tsunamigenic, interplate thrust event (1998.0717 Papua New Guinea) that is considered not to be a tsunami earthquake (Heinrich et al., 2001; Okal, 2003). The low values of $\Theta$ for some strike-slip earthquakes can be attributed to overestimate of $T_0$ for the smaller events, or to anomalously low amplitudes in the $P$ wave train due to the strike-slip radiation pattern. A small error in $T_0$ can give a large error in $\Theta$, since $\Theta$ is proportional to $T_0^{-3}$ (Equation 8). Similarly, the low value $\Theta = -6.0$ for a down-dip earthquake (W; 2005.09.09 New Ireland) can be attributed to overestimate of $T_0$ for this event due to anomalously high-frequency signal in
the depth phases $pP$ and $sP$. The low value $\Theta = -5.9$ for 2003.09.27 Siberia, a continental, strike-slip event, is difficult to explain, but appears related to an excessively long coda in the high frequency seismograms used to estimate $T_0$.

**Identification of interplate thrust, tsunami and other event types**

Even without the moment scaling $M_{wpd}$ provides a closer match to $M_w^{CMT}$ magnitude, including for larger interplate thrust events and tsunami earthquakes, than do most other procedures for rapid magnitude estimation (standard deviation of $\sigma=0.16$; cf. Figures 1 and 4, Table 1). And a “raw” $M_{wpd}$ given by direct application of Equation (3) without any corrections for event type (e.g., no crustal correction for shallow continental events, no correction for radiation pattern for strike-slip events) still matches $M_w^{CMT}$ with $\sigma=0.16$. However, to obtain the best match of $M_{wpd}$ to $M_w^{CMT}$ requires identification of interplate thrust and tsunami earthquakes for application of the moment scaling, and further classification by type, e.g., as continental, oceanic, strike-slip, or deep, for application of corrections for PREM properties at the source depth and for radiation pattern. In practice, information on the location, tectonic setting and likely focal mechanism of an event will usually be available before the duration-amplitude analysis is completed, thus likely interplate thrust and tsunami events, the event type and the approximate source depth can be identified rapidly. The correction for PREM properties at the source depth relative to average, upper-mantle properties is only significant (i.e., giving a magnitude change $\delta M > 0.15$) for events deeper than 220 km. The corrections for continental type, and strike-slip mechanisms are important for oceanic, strike-slip events ($\delta M \approx 0.13$) and continental, reverse and normal faulting events ($\delta M \approx -0.15$), but not for continental, strike-slip events for which these two corrections approximately cancel. However, not all events can be easily classified within minutes after OT. For example, the 12 September 2007, 23:49, M7.9 Indonesia earthquake, is classified here as a downdip event (based on the epicentral location and the CMT centroid depth of 44 km, Table 1) giving $M_{wpd} = 7.9$; but the epicentral location and and shallow initial depth estimate for this event could imply that it is an interplate thrust event, in which case the amplitude-duration moment scaling should be applied, giving $M_{wpd} = 8.2$.

**Application at local and regional distances**

It is likely that the duration-amplitude methodology can be applied at local and regional distances, i.e. $\text{GCD} < 30^\circ$, thus reducing the time delay after OT for obtaining event size estimates. The main difficulty in applying the methodology for seismograms recorded at $\text{GCD} < 30^\circ$ is that significant $S$ signal may remain on the high-frequency, $P$-wave seismograms used for determination of the duration, $T_0$, which complicates the analysis of larger events. In this case, the direct $P$-wave radiation can often be isolated by applying the narrow-band, Gaussian filtering at higher frequencies (e.g., 5-20 Hz), but this requires that high dynamic-range, high sample-rate data is available.

**Physical implications of moment scaling**

The increasing underestimate of $M_w^{CMT}$ with increasing magnitude by unscaled $M_{wpd}$ for interplate thrust (and possibly tsunami) earthquakes (Figure 4) and the consequent need for a moment scaling (Equation 5a) may have important physical implications. The increasing underestimate of $M_w^{CMT}$ is probably not due to station site or path effects, since then it would occur for all event types, and it is probably not a direct effect of the source mechanism radiation pattern, since then it would not vary with event size. In addition, examination of $M_{wpd}$ estimates obtained with different long-period cutoffs (Appendix C) indicates that the
increasing underestimate of $M_{w,CMT}$ is not due to magnitude saturation. Thus the increasing underestimate of $M_{w,CMT}$ may be associated with near-source, dynamic phenomena unique to larger interplate thrust (and possibly tsunami) earthquakes, events which occur at shallow depths. The form of the moment scaling, Equation (5a), suggests a deficiency that increases with event size in the amplitude of far-field, radiated $P$-waves relative to the amplitudes expected from the CMT results.

The destructive interference of $pP$ or $sP$ waves with direct, down-going $P$ waves is an often cited explanation for reduced, far-field $P$ amplitudes, but this is a kinematic mechanism which must be cast into a dynamic framework for large, shallow earthquakes where the interference would occur within the rupture volume and simultaneous with rupture. The deficiency in amplitude must therefore be associated with a near-field mechanism which reduces the radiated kinetic energy while maintaining the seismic energy balance. A candidate mechanism would be excessive dissipation of the strain energy released during faulting by fracturing and frictional processes on or near the fault, alimented by complex wave interactions around the rupturing fault that provide additional dynamic strain energy. These interactions would include waves reflected, generated or trapped near the free surface, such as the near-field analogues of $pP$ and $sP$, and these waves could interfere destructively with the fault displacements that produce far-field $P$-waves, reducing the amplitude of these waves. We can then hypothesise a transfer of kinetic energy along strike and in the direction of rupture (for long thrust faults) by waves from earlier rupture that have interacted with the free surface, producing dynamic stress loading across the fault at the rupture front and augmenting the loading due to nearby fault displacements.

Such dynamic loading near the rupture front could raise the shear stress above the failure yield stress (e.g., Scholz, 2002), decrease the normal stress and thus decrease the effective yield stress (e.g., Oglesby et al., 2000), or drive rupture in zones with a velocity-strengthening friction behavior (e.g., Scholz, 1998). In all these cases, increased fracture, rupture and slip would be induced at the rupture front, including on parts of the fault for which the initial shear stress was much less than the static yield stress, or which have velocity-strengthening behavior, likely in the shallower, up-dip parts of subduction thrusts (e.g., Scholz, 1998). Thus the moment scaling could be a manifestation of a “self-driving” mechanism for large interplate thrust (and possibly tsunami) earthquakes in which an anomalously large proportion of the energy released during rupture is re-absorbed locally to further drive the rupture, and thus to make the earthquake large.

Acknowledgements

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References


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Table 1

Events used in this study and duration-amplitude results

<table>
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<tr>
<th>Origin time</th>
<th>Event</th>
<th>Type*</th>
<th>latitude</th>
<th>longitude</th>
<th>depth (km)</th>
<th>NEIC</th>
<th>CMT</th>
<th>this study</th>
<th>this study, duration-amplitude results</th>
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<td>7.8</td>
<td>7.9</td>
<td>7.9 (2007)</td>
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</tbody>
</table>

* Earthquake type: 1 - intraplate thrust; 2 - tsunami earthquake; W - dominantly P; 3 - dominantly S; 4 - strike-slip event; 5 - reverse faulting; 6 - normal faulting; 7 - strike-slip continental; 8 - reverse faulting continental; 9 - normal faulting continental.

1 s (CMT centred time - origin time).

3 magnitud dependent correction of Whitten et al. (2002).

** This study, duration-amplitude regression analysis due to complex nature of source.

*** M - only for 2002.02.05 Sanriku-Andaman from Tani et al. (2005).
Figure 1 a)

Figure 1 b)
Moment magnitudes from rapid analysis methods using seismic $P$-waves compared to CMT magnitude $M_w^{\text{CMT}}$ for the studied events (Table 1, Figure 2). a) $M_w^{\text{NEIC}}$ from the NEIC Fast Moment Tensor procedure (Sipkin, 1994; http://earthquake.usgs.gov); b) $M_{wp}$ from this study, determined following the procedure described by Tsuboi (2000), Hirshorn (2006) and Lomax et al. (2007); c) $M_{wp}$ from this study with magnitude dependent correction of Whitmore et al. (2002). Event symbols are: interplate thrust events (blue inverted triangles); tsunami earthquakes (red squares); other event types (green diamonds). In this and the following figures the value $M_w^{\text{CMT}}=9.3$ for 2004.12.26 Sumatra-Andaman is from Tsai et al. (2005).

Figure 1

Moment magnitudes from rapid analysis methods using seismic $P$-waves compared to CMT magnitude $M_w^{\text{CMT}}$ for the studied events (Table 1, Figure 2). a) $M_w^{\text{NEIC}}$ from the NEIC Fast Moment Tensor procedure (Sipkin, 1994; http://earthquake.usgs.gov); b) $M_{wp}$ from this study, determined following the procedure described by Tsuboi (2000), Hirshorn (2006) and Lomax et al. (2007); c) $M_{wp}$ from this study with magnitude dependent correction of Whitmore et al. (2002). Event symbols are: interplate thrust events (blue inverted triangles); tsunami earthquakes (red squares); other event types (green diamonds). In this and the following figures the value $M_w^{\text{CMT}}=9.3$ for 2004.12.26 Sumatra-Andaman is from Tsai et al. (2005).
Figure 2

World map showing earthquakes used in this study (c.f. Table 1). Symbols show earthquake type: I - interplate thrust (blue inverted triangles); T - tsunami earthquake (red squares); W – downdip and P – intraplate (light blue triangles); D – deep (green triangles); So - strike-slip oceanic, Ro – reverse-faulting oceanic and No – normal-faulting oceanic (magenta diamonds); S - strike-slip continental, R - reverse-faulting continental and N - normal-faulting continental (yellow diamonds); hybrid events (white diamonds). Base map from NGDC (2006); plate boundaries (magenta lines) from Coffin et al. (1998).
Figure 3

Duration-amplitude processing steps for the 12 September 2007, M8.4 Sumatra earthquake recorded at station IU:KBL at 49º GCD to the northwest of the event. Trace (0): raw, velocity seismogram; Trace (1): 1.5 Hz, Gaussian-filtered seismogram; Trace (2): smoothed, velocity-squared envelope; Trace (3): amplitude-corrected, ground-displacement seismogram; Trace (4): integral of trace (3) over the source duration using Equation 3, note that the integral only accumulates for positive values of displacement in trace (3). P and S indicate the PREM_NC predicted arrival times for the first arriving, P and S waves from the hypocentre.
9, 8, 5 and 2 indicate the times at which the envelope function, trace (2), last drops below 90% ($T^{90}$), 80% ($T^{80}$), 50% ($T^{50}$) and 20% ($T^{20}$) of its peak value, respectively; To indicates the estimated apparent duration, $T_0$, for this station. See Appendix B for more details.
Figure 4

Results for duration-amplitude magnitude $M_{wpd}$ with no moment scaling for interplate thrust or tsunami events (i.e., application of Equation 3) for the studied events (Table 1). a) $M_{wpd}$ compared to CMT magnitude $M_w^{CMT}$, b) $\Delta M = M_{wpd} - M_w^{CMT}$ compared to $M_w^{CMT}$. Material properties at the source are corrected to correspond to the PREM_NC model values at the
CMT centroid depth. The regression against $M_0^{CMT}$ to determine $k$ in Equation (3) is performed excluding interplate thrust and tsunami events and 2002.11.03 Alaska (labelled RS in plots) which has a poor $T_0$ estimate due to exceptional source complexity (e.g., Fuis and Wald, 2003). Event symbols and labels as in Figure 1.
Figure 5

Results for duration-amplitude magnitude $M_{wpd}$ corrected with moment scaling for interplate thrust and tsunami events (i.e., application of Equations 3 and 5a or 5b) for the studied events.
(Table 1). a) $M_{wpd}$ compared to CMT magnitude $M_w^{CMT}$, b) $\Delta M = M_{wpd} - M_w^{CMT}$ compared to $M_w^{CMT}$. Material properties at the source are corrected to correspond to the PREM_NC values at the CMT centroid depth. Event symbols and labels as in Figure 1.
Figure 6

$\Theta$ values from application of Equation (8) to duration-amplitude results with moment scaling for interplate thrust and tsunami events (i.e., application of Equations 3 and 5a or 5b) for the studied events (Table 1). $\Theta$ values are plotted against CMT magnitude $M_{w}^{CMT}$. Lines of constant $\Theta$ are shown for $\Theta = -4.9$, the expected value for all earthquakes, and $\Theta = -5.5$, below which indicates a possible tsunami earthquake. Event symbols and labels as in Figure 1.
Appendix A - Far-field estimation of seismic moment from $P$ waveforms

Following Aki and Richards (1980), Boatwright and Choy (1986), Tsuboi et al. (1995) and Kanamori and Rivera (2004), if $u(t)$ is the amplitude corrected, far-field, $P$-displacement for an earthquake source of duration $T_0$, then a theoretical expression for scalar seismic moment, $M_0$, is,

$$M_0 = C_M \int_{t_p}^{t_p+T_0} u(t) \, dt.$$  \hspace{1cm} (A1)

In the above expression $t_p$ is the $P$ arrival time and $u(t)$ is corrected for geometrical spreading and attenuation. $C_M = 4\pi \rho_s^{1/2} \alpha_s^{5/2} \alpha_r^{1/2} F f_s$, where $\rho$ and $\alpha$ are the density and $P$ wave speed, respectively, at the source $s$ or the recording station $r$, and $F$ and $f_s$ are corrections for radiation pattern and free-surface amplification, respectively.

In this study we use the 1-D, spherical, PREM model (Dziewonski and Anderson, 1981) without a crust (PREM_NC) for calculation of $C_M$ and for amplitude correction of the displacement waveforms for attenuation and geometrical spreading. In PREM_NC the crustal layers are replaced by a layer with the PREM properties of the uppermost mantle. For shallow continental events (Table 1), the effect of the crust on $C_M$ is introduced as a magnitude correction using the PREM properties for the lower crust.

The geometrical spreading is calculated from the spreading of rays between the source and station in the PREM_NC model using a standard expression (e.g., Aki and Richards, 1980, eq. 9.44; Shearer, 1999, eq. 6.23).

The attenuation correction is made in the frequency domain using standard relations (e.g., Shearer, 1999; Lay, 2002),

$$A_{corr}(\omega) = A_0(\omega) e^{-\omega t^*},$$  \hspace{1cm} (A2)

and,

$$t^* = \int_{path} \frac{dt}{Q(r)},$$  \hspace{1cm} (A3)

where $A_0(\omega)$ and $A_{corr}(\omega)$ are the Fourier transforms of the initial and attenuation corrected displacements, and the integral in Equation (A3) is taken using the source-station ray path and corresponding $Q$ values from the PREM_NC model.

If the integral in Equation A1 includes all of the $P$ wave group ($P$, $pP$ and $sP$) then the correction to displacement for radiation pattern is given by a factor $F = \sqrt{\langle F^p \rangle^2 / \langle F^{pP} \rangle^2}$ where $\langle F^p \rangle^2 = 4/15$ (e.g., Boatwright and Choy, 1986) is the mean square radiation coefficient for $P$ waves and $F^{pP}$ is a generalized radiation pattern coefficient for the $P$ wave group. For observations at teleseismic distances Newman and Okal (1998) suggest a constant value $F^{pP} = 1$ for the generalized radiation coefficient which is appropriate for dip-slip faulting but considered too high by as much as a factor of 4 for strike-slip faulting (Boatwright and Choy, 1986; Choy and Boatwright, 1995). This choice of $F^{pP}$ gives $F = \sqrt{\langle F^p \rangle^2} = \sqrt{(4/15)} \approx 0.52$. However, if the integral in Equation A1 includes only the direct $P$ waves, then $F = \sqrt{\langle 1 / (F^p)^2 \rangle} = \sqrt{15/4} \approx 1.9$ (e.g., Tsuboi et al., 1999). Since in this study we compensate for the presence of non direct $P$ waves by taking the integral in Equation (1) separately over the positive and
negative displacement amplitudes, we use $F = \sqrt{(15/4)}$ for the radiation pattern correction for non-strike slip events. Because of the ambiguity noted above in the radiation coefficients for strike-slip faulting, we determine empirically a magnitude correction for strike-slip events so that their $M_{wpd}$ magnitudes best match $M_w^{CMT}$ on average.

The correction for free-surface amplification at the station site introduces an additional factor of $f_s = 1/2$. Incorporating the corrections for radiation pattern and free-surface amplification in $C_M$, and using PREM NC upper-mantle material properties for the source, $\rho_s = 3.38$ g/cm$^3$, $\alpha_s = 8.10$ km/sec, and PREM upper crust properties for the recording stations, $\rho_r = 2.60$ g/cm$^3$, $\alpha_r = 5.80$ km/sec, gives $C_M = 1.62 \times 10^{19}$ when geometrical spreading is expressed as an equivalent source-station distance in units of km. For shallow continental events, the effect of the crust on $C_M$ gives a magnitude correction ($\delta M = -0.15$) using the PREM properties for the lower crust, $\rho_r = 2.90$ g/cm$^3$, $\alpha_r = 6.80$ km/sec.
Appendix B – Duration-amplitude moment and magnitude calculation

For each earthquake we assume that we have a hypocentre location and predicted P and S travel times from the hypocentre to each recording station. Currently, most real-time monitoring agencies have this information within a few minutes after OT for local and regional events (GCD to stations < ~30°), and within about 10 to 15 minutes of OT for teleseismic events (GCD to stations > ~30°). We also assume that we have available vertical-component, broadband, digital seismograms for about 20 or more stations at 30° to 90° GCD from the source and that these stations are moderately well distributed in distance and azimuth. We exclude from the analysis poor quality seismograms that are noisy, clipped, truncated, or otherwise corrupted.

For the present study we examine a set of recent earthquakes with a large range of magnitudes ($M_{w}^{CMT}$ 6.6 to 9.3) and diverse source types (Table 1). For each event, we obtain from the IRIS Data Management Center a set of broadband vertical (BHZ) component recordings at stations from 30° to 90° GCD from the event. Typically we use about 20 to 50 records, selecting stations well distributed in distance for events which have more than 50 available records. All averages and standard deviations are obtained using robust statistics (i.e., rejection of the upper and lower 10 percentiles of values), typically data from 15 to 45 stations are retained.

Duration determination

We estimate the source duration, $T_0$, for each station using vertical-component seismograms and the following procedure (see also Figure 3), based on that of Lomax (2005) and Lomax et al. (2007): 1) Convert the seismograms from each station to high-frequency records using a narrow-band, Gaussian filter of the form $e^{-\alpha|\frac{f-f_{cent}}{f}|^2}$, where $f$ is frequency, $f_{cent}$ the filter center frequency, and $\alpha$ sets the filter width. Here we use $f_{cent} = 1.5$ Hz and $\alpha = 20.0$; this value of $f_{cent}$ is intermediate between the $f_{cent} = 1.0$ Hz of Lomax (2005) and Lomax et al. (2007) and the 2-4 Hz band-pass filter used by Hara (2007). 2) Convert the high-frequency seismogram to velocity-squared time-series by squaring each of the data values. 3) Smooth the velocity-squared time-series with a 10 sec wide, triangle function to form a station envelope function. 4) Measure the set of time delays after the P time at which the envelope function last drops below 90% ($T_{90}$), 80% ($T_{80}$), 50% ($T_{50}$) and 20% ($T_{20}$) of its peak value. 5) Calculate the apparent source duration, $T_{0}$, for the station using the following algorithm,

$$T_0 = (1 - w)T_{90} + wT_{20},$$

where the weight $w = [(T_{80} + T_{50}) / 2 - 20 \text{ sec}] / 40 \text{ sec}$, with limiting values $0 \leq w \leq 1$.

The form of $w$ and choice of 20% and 90% of the envelope peak value to measure $T_0$ follow from examination of the shape of the summary envelope functions used in this study. In general, the 20% peak value gives better agreement with published results for the larger events (e.g., $T_0 > 100$ sec), while the 90% peak value better results for the smallest events (e.g., $T_0 < 100$ sec), in comparison to twice the CMT centroid minus origin times and other estimates of source duration. The necessity for different treatment of smaller and larger events is due to the longer length of the exponentially decaying, P coda in proportion to the source duration for smaller events than for larger events.
We also calculate an average \( T_0 \) and associated standard deviation for each event by taking the geometric mean and geometric standard deviation of the station \( T_0 \) estimates using robust statistics.

**Duration-amplitude moment and magnitude calculation**

We evaluate the seismic moment, \( M_{0^{pd}} \), for each station using vertical-component seismograms and the following procedure (see also Figure 3): 1) Bandpass from 1 to 200 sec (see Appendix C), remove the instrument response and apply geometrical spreading and attenuation corrections to convert each seismogram to amplitude corrected, ground displacement. 2) Cut each seismogram from 10 seconds before the \( P \)-arrival to the \( P \)-arrival time plus the source duration, \( T_0 \), or to 10 seconds before the \( S \) arrival, whichever is earlier, to obtain \( P \)-wave seismograms. 3) Apply Equations (3) and (5a or 5b) to each \( P \)-wave seismograms to obtain station moment estimates. 4) Multiply the station moment value by a factor \( T_0/t_{S-P} \) if \( T_0 > t_{S-P} \), where \( t_{S-P} \) is the \( S \) arrival time minus the \( P \) arrival time. We calculate an average \( M_{0^{pd}} \) and associated standard deviation for each event by taking the geometric mean and geometric standard deviation of the station moment estimates using robust statistics.

We calculate the duration-amplitude magnitude, \( M_{wpd} \), through application of the standard moment to moment magnitude relation (Hanks and Kanamori, 1979),

\[
M_{wpd} = (\log_{10} M_{0^{pd}} - 9.1)/1.5 ,
\]  

(B2)

where \( M_{0^{pd}} \) has units of N-m.

We include a regression constant, \( k \), in Equation (3) to compensate for the errors and biases in the geometrical spreading and attenuation corrections and in the terms of \( C_M \). We evaluate \( k \) through regression of our \( M_{0^{pd}} \) values for each event against the corresponding CMT moment values, \( M_{0^{CMT}} \), so that the mean of \( \log_{10}(M_{0^{pd}}/M_{0^{CMT}}) \to 0 \), giving \( k \approx 1.1 \). This regression excludes interplate thrust, tsunami and strike-slip events and 3 November 2002 Alaska (labelled RS in plots) which has an unstable \( T_0 \) estimate due to exceptional source complexity (e.g., Fuis and Wald, 2003). We use only interplate thrust and tsunami events to perform a regression to determine the constant \( M_0^{cutoff} \) in Equation (5a), giving \( M_0^{cutoff} \approx 7.5 \times 10^{19} \) N-m (equivalent to \( M_w \approx 7.2 \)), and to determine the optimal value of \( R \) in Equation (4) by minimizing the standard deviation of \( \log_{10}(M_{0^{pd}}/M_{0^{CMT}}) \), giving \( R \approx 0.4 \). The optimal values of \( M_0^{cutoff} \) and \( R \) are sensitive to details of the algorithms used to estimate \( T_0 \) and moment; a change of \pm0.25 in \( R \) gives about half the variance reduction relative to \( R = 0 \) (i.e., no moment scaling) than gives \( R \approx 0.4 \). The empirically determined magnitude correction to account for the radiation pattern of strike-slip events (types S and So in Table 1) has a value of 0.13 magnitude units; this value implies that for strike-slip events an additional factor of about 1.6 is needed in the correction for radiation pattern, \( F \), in Equation (A1).
Appendix C – Dependence of duration-amplitude results on long-period cutoff

The values of moment and of moment magnitude, $M_{wpd}$, for large events obtained with the duration-amplitude procedure depend on the long-period cutoff used when processing the seismograms. Indeed, it is generally accepted that magnitude saturation, regardless of the magnitude estimation technique, is related to the long-period, data cutoff being lower than a corner period above which the displacement spectrum flattens to an amplitude proportional to the static moment (e.g., Stein and Okal, 2006). Magnitude saturation also arises for methods that use a signal duration after the initial $P$ arrival that is shorter than the duration of significant $P$ signal and the source duration (e.g., Granville et al., 2005); the duration-amplitude procedure avoid this problem by explicitly taking into account the source duration.

Figure C1 shows duration-amplitude magnitudes, $M_{wpd}$, with no moment scaling, for the 7 largest and one tsunami earthquake from the studied events, plotted as a function of long-period cutoff, $T_{cutoff}$. With increasing $T_{cutoff}$ to about 50-100 sec there is an increase in magnitude estimates for all events; this increase can be associated with magnitude saturation due to $T_{cutoff}$ being lower than the long-period spectral corner for $P$ waves. However, at around $T_{cutoff} = 100-200$ sec the curves in Figure C1 flatten and the magnitude estimates are nearly independent of $T_{cutoff}$, indicating that the long-period corner for $P$ waves for each event has been reached and that the resulting magnitude estimates should not be saturated. Above around $T_{cutoff} = 200$ sec the magnitude estimates for some events again increase with $T_{cutoff}$; examination of the processed seismograms shows that this increase is primarily an artefact of amplification of long-period noise during the removal of the instrument response. The onset of noise above about 200 sec period is expected since the typical long-period corner is about 120 to 360 sec for the very-broadband instruments providing much of the data used in this study.

These results and Figure C1 indicate that: 1) The optimal long-period cutoff for the studied data set is 100-200 sec. 2) The trend of increasing underestimate of $M_w^{CMT}$ by unscaled $M_{wpd}$ with increasing $M_w^{CMT}$ (Figure 4) cannot be attributed to a standard magnitude saturation problem due to insufficient long-period signal.
Figure C1

Duration-amplitude magnitudes, $M_{wpd}$, with no moment scaling (i.e., application of Equation 3) for the 7 largest and one tsunami earthquake (2006.07.17 Indonesia) from the studied events (Table 1) plotted as a function of long period cutoff used for analysis. The events are identified by their origin dates.