A NEW WAY TO LOOK AT FAULT RUPTURE:
UNDERSTANDING AND CHARACTERIZING EARTHQUAKE PREDOMINANT
PERIOD

by

Erik L. Olson

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A NEW WAY TO LOOK AT FAULT RUPTURE:
UNDERSTANDING AND CHARACTERIZING EARTHQUAKE
PREDOMINANT PERIOD

Approved by:

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Date: __________________________

Richard M. Allen
Assistant Professor of Geophysics
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I would also like to thank my friends and family, both around Madison, and every place I’ve visited during my time at the University of Wisconsin. Without their support both toward accomplishing my research, and taking a break from it, I would not have been able to complete this document. Thank you to everyone who has helped make this a great experience. I would also like to thank you, for showing interest in my work and reading this text.
INTRODUCTION

Earthquakes are a fascinating phenomenon that over the course of millions of years have changed the face of our planet. Earthquakes remain an enigma, defying scientists’ best attempts to characterize their behavior. Until the mid-1900’s we could not even fully characterize the size of an earthquake; only through assessing the destruction wrought by the event could we guess at the size of the force. In the late 1930’s the magnitude scale arose as a new method of understanding the strength of an earthquake [Stein and Wysession, 2003]. Even after the magnitude scale was developed, another 30 years passed before magnitude was tied to earthquake moment, and therefore related to slip on the fault [Stein and Wysession, 2003].

Throughout history, even before the size of an earthquake was well understood, people attempted to predict these events. However, calculating the magnitude of an earthquake was not possible, even during the earthquake, let alone prior to rupture initiation. As rupture theory developed, and more earthquakes were studied, scientists began to postulate that the very nature of rupture prevented any fore-warning of an earthquake, and that an earthquake’s magnitude could only be determined post-rupture.

Several other theories were proposed contradicting the non-deterministic model of earthquake rupture, however they often failed to explain earthquakes in more than one region, or fell into disfavor after additional research weakened supporting arguments. Then, a few years ago, new empirical evidence appeared which hinted at the possibility of a deterministic rupture process, i.e. evidence that the size of an earthquake may be determined prior to rupture termination. Working with a dataset from California, Allen and Kanamori [2003] found that the predominant period of an earthquake scales with the magnitude of the
earthquake. The maximum predominant period of an earthquake, a value calculated from the frequency content of the P-wave, could be measured in 5 seconds, which is faster than a large earthquake takes to rupture. However, the study of Allen and Kanamori [2003] was limited to a small region, and only included 3 events greater than magnitude 6.

Lockman and Allen [In Review] took the analysis of maximum predominant period a step further and added earthquakes from Japan and the Pacific Northwest to the data set. These additional events remained consistent with the original observations that the maximum predominant period of an earthquake is related to the magnitude of the earthquake. Unfortunately the additional data set did not substantially increase the number of large magnitude events, and the source of predominant period remained an enigma.

The study presented in the following chapters was initiated to pick up the next step in the development of our understanding of predominant period. Despite several years of study, and the publication of many papers on the subject, no one has yet pinned down what causes the earthquake predominant period; nor what process occurs during earthquake rupture initiation that affects the event’s final magnitude. This study represents the beginning of this characterization.

References Cited

CHAPTER 1

The Deterministic Nature of Earthquake Rupture

Erik L. Olson
Department of Geology and Geophysics, University of Wisconsin Madison, 1215 W. Dayton St., Madison, WI 53706, USA

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Introduction

Understanding the earthquake rupture process is key to our understanding of fault systems and earthquake hazards. Over the past 15 years multiple hypotheses concerning the nature of fault rupture have been proposed but no unifying theory has emerged \(^1-^{11}\). The conceptual hypothesis most commonly cited is the cascade model for fault rupture \(^2,^{6,7,12}\). In the cascade model, slip initiates on a small fault patch and continues to rupture further across a fault plane as long as the conditions are favorable. Two fundamental implications of this domino-like theory are that small earthquakes begin in the same manner as large earthquakes, and that the rupture process is not deterministic, i.e., the size of the earthquake cannot be determined until the cessation of rupture. Here we show that the frequency content of radiated seismic energy within the first few seconds of rupture scales with the final magnitude of the event. Therefore the magnitude of an earthquake can be estimated before the rupture is complete. This finding implies that the rupture process is to some degree deterministic and has far-reaching implications for the physics of the rupture process.

Rupture dynamics background

In the cascade model faults are divided into patches of varying size and shape. When an earthquake initiates on one patch, slip on this patch can lead to continued slip on the adjacent patches if the rupture energy and the state of stress on adjacent patches are favorable. An earthquake continues spreading from patch to patch until there is insufficient energy to rupture the next patch at which point the rupture stops. Moment is physically related to both rupture area, \(A\), and average slip, \(\bar{D}\), by the relation \(M_o = \mu A \bar{D}\) where \(M_o\) is the seismic moment and \(\mu\) is the shear modulus. Moment magnitude, \(M_w\), scales with the
seismic moment and is therefore similarly related to $A$ and $\overline{D}^{13}$. Given this framework, it is not possible to know the magnitude of an earthquake until the rupture has stopped.

Throughout the last decade the seismological community has debated whether the first few seconds of the P-wave (the first few seconds of radiated energy) provides information about the final magnitude of an earthquake before the rupture is complete $^{3,6,7,10}$. Much of the debate has focused on the time-domain characteristics of the P-wave. However, evidence for a scaling relation between the frequency content of the first few seconds of the P-wave and the final magnitude has also emerged $^{14-17}$. Using an approach similar to Nakamura$^{16}$, Allen and Kanamori$^{14}$ measured the predominant period, $\tau_p$, from the first 4 seconds of the P-wave arrival at multiple seismic stations in southern California. $\tau_p$ is calculated from the velocity waveform using the following recursive relation

$$\tau_i^p = 2\pi \sqrt{X_i/D_i} \quad (1)$$

where

$$X_i = \alpha X_{i-1} + x_i^2, \quad D_i = \alpha D_{i-1} + \left(\frac{dy}{dt}\right)_i \quad (2)$$

$x_i$ is the ground motion recorded at time $i$ and $\alpha$ is a 1 sec smoothing constant$^{14}$. Allen and Kanamori$^{14}$ demonstrated a scaling relation between $\tau_p$ and magnitude $M$ for earthquakes with $M$ of 3.0 to 7.3. In earthquakes with $M < 6$ the total duration of rupture is usually less than 4 sec, therefore, the entire rupture time history is included within the first 4 sec of the P-wave. However, for $M > 6$ earthquakes, the existence of a scaling relation between $\tau_p$ and $M$ would imply that the magnitude of an earthquake has been defined before the rupture terminates and that the rupture process is deterministic. Allen and Kanamori$^{14}$ used
earthquakes from southern California where data for only 3 earthquakes with \( M_w > 6 \) are available.

**Methods and data**

Here, we measure \( \tau_p \) for a much larger number of earthquakes including events from Japan, Taiwan, California and Alaska. The waveform data has been provided by K-net operated by the National Research Institute for Earth Science and Disaster Prevention in Japan, the Taiwan Strong Motion Instrumentation Program of the Central Weather Bureau, Southern California Seismic Network operated by Caltech and the US Geological Survey, and the University of Alaska Geophysical Institute. A total of 71 earthquakes producing 1842 waveforms recorded on both broadband velocity sensors and accelerometers within 100 km of the epicenter are used. There are 24 events with \( M_w \geq 6.0 \), including the \( M_w 7.6 \) Chi-Chi earthquake (Taiwan, 1999) with a rupture duration of \( \sim 30 \) sec, the \( M_w 7.9 \) Denali earthquake (Alaska, 2002) with a rupture duration of \( \sim 70 \) sec, and the \( M_w 8.3 \) Tokachi-oki earthquake (Japan, 2003) with a rupture duration of \( \sim 40 \) sec.

We calculate \( \tau_p \) in a recursive fashion from a vertical velocity timeseries to generate \( \tau_p(t) \) as a function of time. Figure 1 shows the vertical velocity waveform recorded during a \( M 4.6 \) earthquake in southern California and the \( \tau_p \) timeseries derived from it. Figure 2 shows a similar example but for the \( M_w 8.3 \) Tokachi-oki earthquake. In this case only acceleration records are available which have been recursively integrated in a causal fashion to derive the velocity trace from which \( \tau_p(t) \) is derived. We define the parameter \( \tau_p^{\text{max}} \) as the maximum \( \tau_p(t) \) data point between 0.05 and 4.0 sec after the P-wave trigger as shown in
Figures 1 and 2. The time window starts at 0.05 rather than 0.00 sec due to the recursive nature of the $\tau_p(t)$ calculation. Using the maximum value between 0.00 and 4.0 can result in leakage of the frequency content of background noise before the P-wave arrival into the time window after the P-wave.

When $\tau_p^{\text{max}}$ is plotted against $M$ on a log-linear scale, a scaling relation emerges as shown in Figure 3a. The $\tau_p^{\text{max}}$ observations from single waveforms have been averaged for each earthquake using all available data. The best-fit linear relation is

$$\log \tau_p^{\text{max}} = 0.14M - 0.83$$

and the average absolute deviation of the event averages is 0.54 magnitude units. While there is scatter in the individual event observations, the scaling relation is clear, implying that information about the final magnitude of an earthquake is available within 4 sec of rupture initiation irrespective of the total rupture duration.

A second parameter, $\tau_d$, is also measured from the $\tau_p$ timeseries. $\tau_d$ is the delay of the $\tau_p^{\text{max}}$ observation with respect to the P-wave trigger and is therefore in the range of 0.05 to 4 sec (see Figures 1 and 2). Figure 3b plots the event-averaged $\tau_d$ observations versus magnitude which shows a general increase in $\tau_d$ with magnitude. Also indicated on Figure 3b is the typical rupture duration of earthquakes as a function of magnitude. This relation shown is only approximate, and is based on scaling relations between rupture length and magnitude. The rupture length is converted to rupture duration by assuming unilateral rupture and a rupture velocity between 2.4 and 3.0 km/s from the estimates of Somerville et al\textsuperscript{18}. Rupture length may be empirically calculated from earthquake magnitude\textsuperscript{19}, or estimated from the scaling relationship between moment and stress drop. Both methods are
used here to create the range shown in Figure 3b. The actual rupture duration for a given magnitude event can vary by a factor of 2 or 3. Despite the uncertainty in the rupture duration of the earthquakes included in this study, it is clear that for earthquakes with $M > 4.5$ the $\tau_p^{\text{max}}$ observation is made before the rupture has ceased. While up to 4 sec of data are used to determine $\tau_p^{\text{max}}$, the average time window of the P-wave required to determine $\tau_p^{\text{max}}$ is less than 2 sec for almost all earthquakes in our study.

Using published moment rate functions for the $M > 6.0$ events in our study we can estimate the amount of moment release at time $\tau_d$. Rupture directivity effects of finite faults mean that the first 2 sec of the P-wave at a given station does not sample exactly 2 sec of the rupture. However, because our $\tau_d$ measurement is event-averaged we can use it as an approximate estimate of the rupture duration within which $\tau_p^{\text{max}}$ information is available. The amount of seismic moment released within $\tau_d$ increases with magnitude, which is not surprising given that $\tau_d$ also increases with magnitude. But the percentage of moment released is always small (less than 40% for $M > 6$ events) and decreases with magnitude. In the case of the largest earthquakes, Chi-Chi, Tokachi-oki, and Denali, the percentage of moment released at $\tau_d$ is less than 2%.

**Discussion and new hypotheses**

All of these observations challenge the cascade model for fault rupture. They imply that the beginnings of small and large earthquakes are different, and that the rupture process is at least partly deterministic, i.e. the final magnitude of an event is to some degree controlled by processes within the first few seconds of rupture. While there is scatter in the
data, most events fall within ±1 magnitude unit of the best-fit line. The scatter could be due to source processes and/or local site and measurement errors. If the scatter is non-source related, then removal or correction for site and path effects could reduce the scatter in the data points of Figure 3a to a single line, implying that the final magnitude of an earthquake is entirely determined within the first 4 sec of rupture. Variability in the quality of the $\tau_p^{\text{max}}$ observations at different stations has already been observed across the seismic network in southern California, indicating that site effects do play a role\textsuperscript{20}. Nevertheless, it seems unlikely that all the scatter is due only to site effects. Instead, source related processes including rupture behavior, stress heterogeneity and other on-fault variability probably also play a role.

The $\tau_p^{\text{max}}$ observations suggest that the final magnitude of an earthquake is partially controlled by the initiation process within the first few seconds of rupture, and partially by the physical state of the surrounding fault plane. The role played by the initiation process can be understood by considering the energy balance of fault rupture. A rupture can only propagate when the available energy is sufficient to supply the necessary fracture energy\textsuperscript{21,22}. When a propagating fracture encounters a small stress-drop patch, the total energy in the system will begin to decrease. Depending on the size of the patch, rupture may terminate. The total rupture energy available increases with the amount of slip, so a large-slip rupture will propagate further across a heterogeneous fault plane. Therefore, if the rupture pulse initiates with large slip, it is more likely to evolve into a large earthquake. This explanation is also consistent with a recent study of the location of earthquake hypocenter with respect to
the slip distribution. Mai et al. \textsuperscript{23} show that hypocenters are preferentially located within or close to regions of large slip.

The $\tau_p^{\text{max}}$ scaling relation provides new constraints on the physics of the rupture process. Whereas the cascade model suggests that the magnitude of an earthquake is dependent on the state of stress across the fault plane and that the nucleation of all earthquakes is identical, our new observations demonstrate the significance of the rupture initiation process within the first few seconds of an event. Understanding the physics of this process will enable us to predict the magnitude of earthquakes without the need for accurate knowledge of the surrounding state of stress across a fault plane.

\textbf{References Cited}


Acknowledgements

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Figure 1. Example waveform and $\tau_{p}^{\text{max}}$ calculation for a $M_{4.6}$ earthquake in southern California recorded at station GSC 74 km from the epicenter. a, The raw vertical component waveform recorded by a broadband velocity sensor. b, Ten seconds of the velocity waveform after low-pass filtering at 3 Hz. The P-wave trigger time is shown by the vertical line at 13.01 sec. c, $\tau_{p}(t)$ trace calculated in a recursive fashion from the waveform in b showing the
change in the frequency content from the pre-trigger noise to the post-trigger P-wave. The $\tau_p^{\text{max}}$ observation is circled (equal to 0.86 sec in this case), $\tau_d$ is the delay of $\tau_p^{\text{max}}$ with respect to the trigger (0.43 sec in this case).

Figure 2. Example waveform and $\tau_p^{\text{max}}$ calculation for the $M_w$ 8.3 Tokachi-oki earthquake recorded at station HKD112 71 km from the epicenter. a, The raw vertical component waveform recorded on an accelerometer. b, Ten seconds of the raw acceleration waveform.
The P-wave trigger is shown by the vertical line at 35.41 sec. c, Ten seconds of the velocity waveform determined from the acceleration recording using recursive relations only. It has also been low-pass filtered at 3 Hz. d, $\tau_p(t)$ trace calculated in a recursive fashion from the waveform in c. The $\tau_p^{\text{max}}$ observation is circled ($\tau_p^{\text{max}} = 1.62$ sec, $\tau_d = 1.49$ sec), it has a longer period and is observed later than the example in Figure 1c due to the larger magnitude of the earthquake.

\begin{figure}[h]
\centering
\includegraphics[width=\textwidth]{figure3.png}
\caption{The relation between $\tau_p^{\text{max}}$, $\tau_d$ and magnitude. a, The scaling relation between event averaged $\tau_p^{\text{max}}$ and magnitude for earthquakes in southern California (triangles), Japan (circles), Taiwan (squares) and Alaska (star). The best fit line is also shown; the average absolute deviation of the observations is 0.54 magnitude units. b, $\tau_d$ plotted against magnitude showing the general increase in the time required to make the $\tau_p^{\text{max}}$ observation with increasing magnitude. The symbols are the same as in a. The grey bar shows the} \end{figure}
approximate rupture duration as a function of magnitude indicating that the $\tau_p^{\text{max}}$ observation is made before rupture ceases for all $M > 4.5$ earthquakes.

### Table 1

Earthquakes included in this study

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<th>Date</th>
<th>Time (UTC)</th>
<th>Magnitude</th>
<th>Latitude</th>
<th>Longitude</th>
<th>$\tau_p^{\text{max}}$</th>
<th>$\tau_d$</th>
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CHAPTER 2

Fundamental Controls on the Behavior of Earthquake Predominant Period

Erik L. Olson
Department of Geology and Geophysics, University of Wisconsin Madison, 1215 W. Dayton St., Madison, WI 53706, USA
Abstract

Earthquake predominant period is a new characteristic of earthquakes that scales with earthquake magnitude. Previous studies describing predominant period (\( \tau_p^{\max} \)) focused on the relationship between \( \tau_p^{\max} \) and magnitude, but did not examine factors that potentially control the behavior of \( \tau_p^{\max} \). The nature of earthquake rupture, the path of emitted seismic waves, and site effects around a seismometer may all influence the recorded \( \tau_p^{\max} \) values for an earthquake. These effects may be a source of noise, and scatter, in the \( \tau_p^{\max} \) trend, limiting our understanding of the relationship between \( \tau_p^{\max} \) and magnitude. The island of Taiwan represents an ideal location to study effects on \( \tau_p^{\max} \) due to the dense station coverage, and large number of earthquakes in the region. Through examining the spatial relationships of \( \tau_p^{\max} \) recorded for several earthquakes around Taiwan, I demonstrate that local site effects, such as geology and structure may influence the average \( \tau_p^{\max} \) value for an earthquake. These spatial relationships also suggest that the directive nature of rupture influences the average \( \tau_p^{\max} \) for an event. One way to remove the effects mentioned above is through station correction factors obtained through data inversion. Unfortunately, the data set for Taiwan is currently too sparse for the derivation of reliable station corrections. However, the addition of more earthquakes to the data set will help constrain correction factors needed in Taiwan, and help the analysis of other global data sets.

Introduction

Earthquake predominant period is a versatile new characteristic that can be measured in the first four seconds of earthquake rupture. Research demonstrates that the maximum
predominant period measured in the first four seconds of an earthquake ($\tau_p^{\text{max}}$) scales with the magnitude of an earthquake (figure 1), and that constructing an early warning system that uses $\tau_p^{\text{max}}$ can help mitigate some seismic hazard [Allen and Kanamori, 2003; Lockman and Allen, In Review; Lockman and Allen, In Review; Olson and Allen, In Review]. However $\tau_p^{\text{max}}$ remains an empirical measurement that lacks a fundamental explanation in relation to the rupture that creates the signal. Additionally, while factors affecting the seismic waveform that carries the $\tau_p^{\text{max}}$ signal have been studied for several decades, factors that affect $\tau_p^{\text{max}}$ remain unknown.

Site effects, ray path geometry and geology, and rupture directivity are just a few of the factors known to effect the seismic signal recorded at a station. All of these factors were studied at locations around the world, and geoscientists derived correction factors to compensate for many effects of these factors. $\tau_p^{\text{max}}$ however, is a new characteristic, so the effects of station location, ray path geometry, and rupture directivity are all unknown.

**Study Region**

The island of Taiwan represents an ideal location to study the fundamental controls on the behavior of $\tau_p^{\text{max}}$. Due to numerous large earthquakes that occur in the region, local geoscientists installed several hundred seismic stations to study the events. These stations are well distributed across the island covering most of the region with the exception of the central mountains (Figure 2). These stations have recorded hundreds of earthquakes, including 8 events greater than magnitude 5.9 since 1999, on or within 50 km of the island.
Eventually a global study of controls on $\tau_p^{\text{max}}$ will be required, but Taiwan represents a good area to initiate this study. The Taiwan region has several strengths, mentioned above, but also has a few weaknesses when compared with other seismic regions around the globe. Only a few large magnitude earthquakes occurred in Taiwan over the past several years, and two of the larger events are aftershocks of the Chi Chi earthquake. Additionally, none of these large earthquakes represent a well known directive earthquake, such as Hector Mine (October, 1999, California), and Denali (November 2002, Alaska). A well studied directive rupture would allow a detailed examination of the directivity effect on $\tau_p^{\text{max}}$.

Data

$\tau_p^{\text{max}}$ is related to earthquake magnitude (figure 1)[Allen and Kanamori, 2003; Lockman and Allen, In Review; Lockman and Allen, in review; Olson and Allen, In Review]. However the relation presented by Olson and Allen [In Review] shows a significant amount of scatter, allowing $\tau_p^{\text{max}}$ to only predict the magnitude within a ±1 magnitude unit scatter (figure 1). They hypothesized the two most probable sources of scatter are site effects and source effects. Source effects are related to the fault rupture, and may be caused by factors such as fault plane asperities, in-fault materials such as gouge or clay, and the nature of patch rupture. Although these effects can be studied in greater detail, no amount of research will allow correction for the scatter they cause, as each earthquake ruptures a fault in a new way.

Site effects, such as station geology, station noise levels, and perhaps even nearby structures such as basins could potentially be corrected for, therefore reducing the scatter of $\tau_p^{\text{max}}$. To analyze these effects, we gathered data from 8 large (greater than magnitude 5.9)
earthquakes in the Taiwan region. These 8 earthquakes were recorded at over 300 stations across the island, and provide a large data set to work with for examining station effects (Figure 2).

**Analysis**

One qualitative way to examine site effects that control \( \tau_p^{\text{max}} \) measurements is to plot \( \tau_p^{\text{max}} \) at each station that recorded one particular earthquake on a regional map (Figures 3-8). Examining these maps provides a first order estimate of numerous possible site effects. A cluster of stations recording anomalously high or low \( \tau_p^{\text{max}} \) values may be indicative of a regional effect, such as local geology or structure. If the cluster is within a relatively uniform region, geologically, then the cluster of anomalous stations may be indicative of path effects, that only effect a small azimuth range away from the main shock. Two events that occur in the central portion of the island do not produce regular deviations in \( \tau_p^{\text{max}} \) values (Figures 3 and 4). While one region may record anomalously low \( \tau_p^{\text{max}} \) values in one earthquake, the next earthquake recorded in the region displays no such trend.

The events mapped in figures 4 and 5 both show patterns for \( \tau_p^{\text{max}} \) deviation. In both cases, the region of Taiwan centered around 25° latitude and 121.2° longitude contains stations that consistently measure \( \tau_p^{\text{max}} \) values higher than the mean. This phenomenon may be due to regional structures, or the type of rock in the area. These stations are largely confined to a basin north of the central mountain range. However, stations located in other basins throughout the island do not show similar behavior (e.g. southwestern stations in
figure 4), so a more detailed study of the regional structure effects is required. Studying the
\( \tau_p^{\text{max}} \) deviation maps for other events triggering these northwestern stations would also help
find the source of the anomaly in this region. If maps of \( \tau_p^{\text{max}} \) from events occurring north or
west of this anomalous region, do not show a similar trend to figures 5 and 6 then perhaps the
anomaly is caused by a structure between the events off the east coast of Taiwan, and the
stations in the northwest.

The station anomaly pattern in figure 7 resembles the pattern that would be produced
by directive earthquake rupture. Stations south of the event record a higher than average
\( \tau_p^{\text{max}} \), while stations north of the event record a lower than average \( \tau_p^{\text{max}} \). Many of the
stations in figure 7 also recorded the events shown in figures 3-6, and for those events these
stations do not show a similar pattern. Therefore, only the earthquake mapped in figure 7 is
responsible for the observed pattern, rather than regional geology or structure. If this event
ruptured along a fault from south to north, the Doppler effect would cause stations south of
the earthquake to measure lower than average frequencies (higher periods), and the stations
to the north would measure higher than average frequencies (lower periods). The anomaly
pattern in figure 7 may be evidence of this Doppler shift. However, due in part to the small
size of this earthquake, researchers have not determined whether or not this earthquake
ruptured in a unilateral manner. So the apparent directivity signal of figure 7 may just be a
random anomaly pattern.

A more quantitative way to evaluate station effects controlling \( \tau_p^{\text{max}} \) is to invert for
station corrections. For example, if one station consistently recorded anomalously high \( \tau_p^{\text{max}} \)
values, the correction for that station would lower the $\tau_p^{\text{max}}$ value, bringing the station closer to the best fit line of Olson and Allen [In Review]. Here we invert for both a station correction and the best fit line between all of the data, using the same data set as Olson and Allen (submitted). All of the data are used to constrain the best fit line, while only data from Taiwan are used to analyze station corrections. The only constraint on a station correction comes from the number of events recorded at that station. For this reason, we only inverted for station corrections at stations that recorded at least 3 events. We also tried more constrained data sets, inverting for corrections at stations with at least 4, 5, and 6 events being recorded. Ideally every station used in the study would record every event, allowing for the maximum constraint on $\tau_p^{\text{max}}$, however we limited our study to only include stations within 100km of the earthquake epicenter. Therefore, very few stations recorded all 8 events, and only when we step back to a requirement of 3 or more events do we achieve a suitable percentage of total stations.

The inversion method used is a SVD pseudo-inverse. The inversion solves for the $a$ and $b$ values of the best fit line given by $y = ax + b$, and the correction factor for each Taiwan station recording at least the minimum number of events. Table 1 shows the number of data, and number of model parameters inverted for with each different minimum number of events. Using the standard deviation of the inversion residuals, we developed a 95% confidence interval for each one of the model parameters. Comparing the 95% confidence interval with the size of the station correction demonstrates that this problem is poorly constrained. The 95% confidence uncertainty range is larger than the correction factor in almost all cases (Table 2). However, the best fit line parameters that were analyzed, $a$ and $b$,
have very small uncertainty relative to their size, suggesting these two model parameters are well constrained. To achieve constraint on the station corrections similar to those of a and b, a much larger data set is required for the Taiwan region. Increasing from at least 3 events to at least 6 minimum events only slightly improves the constraint, so several more earthquakes are required, perhaps as many as twenty or thirty more to constrain the station correction factors. Figures 9 and 10 are maps of the correction factors calculated from the inversions requiring a minimum of 3 events, and a minimum of 4 events.

Another shortcoming of the inversion is the trade-off allowed between model parameters. The off diagonal elements of the correlation matrix are large, suggesting that many of these corrections trade off. Again, a larger number of events may help constrain the problem, but there will always be a certain amount of trade-off in these values.

**Discussion**

The maps of $\tau_p^{\text{max}}$ deviation from event averages (figures 3-8) highlight many of the station effects that may be influencing $\tau_p^{\text{max}}$. Stations in certain areas of Taiwan appear to consistently measure anomalous $\tau_p^{\text{max}}$ values (figures 5 and 6), and one anomaly pattern suggests the possibility of rupture directivity effecting $\tau_p^{\text{max}}$ (figure 7). However, a more detailed study, including several more moderate to large earthquakes (M>5.0) is required before any of these patterns can be thoroughly understood. With only two events being recorded at the northwestern Taiwan stations (figures 5 and 6) the evidence for a geologically sourced $\tau_p^{\text{max}}$ anomaly is weak. To study the effect of rupture directivity on $\tau_p^{\text{max}}$, the event shown in figure 7 is not the best starting point. Other regions of the world, such as Turkey,
and California, experienced several large earthquakes recently which are known to have ruptured in one direction. These well studied events would be the best starting place for a study in rupture directivity.

Station correction factors derived to identify and correct for any station showing consistent deviation from average $\tau_p^{\text{max}}$ values are too poorly constrained to significantly reduce scatter in the $\tau_p^{\text{max}}$ - magnitude trend presented by Olson and Allen. Similar to the maps discussed above, expanding the data set by adding more moderate sized earthquakes may help constrain these correction factors.

The preliminary evidence discussed above, both from the station correction factors and maps of $\tau_p^{\text{max}}$ suggests that $\tau_p^{\text{max}}$ is influenced by station effects. Eventually correcting for these effects through the use of a larger data set will potentially reduce the scatter in $\tau_p^{\text{max}}$ period relationship, allowing a more accurate rapid estimation of magnitude. These corrections could eventually lead to a more complete understanding of rupture initiation.

References Cited


Acknowledgements
I would like to thank Richard Allen, Cliff Thurber, Michael Brudzinski, and Heather DeShon for many helpful discussions that significantly improved this paper. Funding for this work was provided by USGS/NEHRP awards 03HQGR0043 and 05HQGR0074, and by the University of Wisconsin – Madison.
Figure 1. The scaling relationship between event averaged $\tau_p^{\text{max}}$ and earthquake moment magnitude (from Olson and Allen [In Review]). Earthquakes from California (triangles), Japan (circles), Taiwan (squares), and Alaska (star) all follow a similar relationship between magnitude and period. The scatter is approximately $\pm1$ magnitude unit.
Figure 2. Map of Taiwan, and all of the seismic stations currently installed on the island.
Figure 3. Map of stations within 100km of the M7.6 Chi Chi earthquake that occurred September 20th 1999. The color scale illustrates log10($\tau_p^{\text{max}}$) deviation from the mean $\tau_p^{\text{max}}$ for this earthquake.
Figure 4. Map of stations within 100km of a M6.4 earthquake that occurred on June 10th 2000. The color scale illustrates $\log_{10}(\tau_p^{\text{max}})$ deviation from the mean $\tau_p^{\text{max}}$ for this earthquake.
Figure 5. Map of stations within 100km of a M5.9 earthquake that occurred on June 14th 2001. The color scale illustrates $\log_{10}(\tau_p^{\text{max}})$ deviation from the mean $\tau_p^{\text{max}}$ for this earthquake.
Figure 6. Map of stations within 100km of a M6.1 earthquake that occurred on May 15th 2002. The color scale illustrates $\log_{10}(\tau_p^{max})$ deviation from the mean $\tau_p^{max}$ for this earthquake.
Figure 7. Map of stations within 100km of a M5.9 earthquake that occurred on June 10th 2003. The color scale illustrates $\log_{10}(\tau_p^{\text{max}})$ deviation from the mean $\tau_p^{\text{max}}$ for this earthquake.
Figure 8. Map of stations within 100km of a M7.1 earthquake that occurred on March 31st 2002. The color scale illustrates log10($\tau_p^{\text{max}}$) deviation from the mean $\tau_p^{\text{max}}$ for this earthquake.
Figure 9. Map of the station correction factors for stations recording 3 or more earthquakes. The color scale illustrates $\log_{10}(\tau_p^{\max})$ deviation from the mean $\tau_p^{\max}$ for this earthquake. Stars represent the 8 Mw $> 5.9$ earthquakes since 1999.
Figure 10. Map of the station correction factors for stations recording 4 or more earthquakes.

The color scale illustrates $\log_{10}(\tau_p^{\max})$ deviation from the mean $\tau_p^{\max}$ for this earthquake.

Stars represent the 8 Mw $> 5.9$ earthquakes since 1999.
Table 1: Number of data based on minimum number of events

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<tr>
<td>4</td>
<td>1437</td>
<td>536</td>
<td>111</td>
</tr>
<tr>
<td>5</td>
<td>1153</td>
<td>252</td>
<td>40</td>
</tr>
<tr>
<td>6</td>
<td>1088</td>
<td>187</td>
<td>28</td>
</tr>
</tbody>
</table>

Table 2: Parameters derived from inversion

<table>
<thead>
<tr>
<th>Minimum number of Events</th>
<th>A log10(Tp)/mag</th>
<th>B log10(Tp)</th>
<th>Mean absolute station correction log10(Tp)</th>
<th>Mean absolute confidence interval log10(Tp)</th>
<th>Standard deviation log10(Tp)</th>
</tr>
</thead>
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<tr>
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<td>-0.6898</td>
<td>0.1461</td>
<td>0.1968</td>
<td>0.1742</td>
</tr>
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<td>-0.6956</td>
<td>0.1061</td>
<td>0.1731</td>
<td>0.1755</td>
</tr>
<tr>
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<td>-0.7144</td>
<td>0.0893</td>
<td>0.1582</td>
<td>0.172</td>
</tr>
<tr>
<td>5</td>
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<td>-0.7549</td>
<td>0.0671</td>
<td>0.1333</td>
<td>0.1669</td>
</tr>
<tr>
<td>6</td>
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<td>-0.7633</td>
<td>0.0688</td>
<td>0.1251</td>
<td>0.1651</td>
</tr>
</tbody>
</table>
CONCLUSIONS AND FUTURE RESEARCH

\( \tau_{p}^{\text{max}} \) scales with earthquake magnitude for any earthquake between magnitude 3.0 and magnitude 8.3. The \( \tau_{p}^{\text{max}} \) of an earthquake can be measured within the first four seconds of rupture, even if the rupture has not ceased. Previous rupture models that suggested earthquake rupture behaved in a non-deterministic manner cannot explain this new observation, and must be modified, or replaced by a new theory of fault rupture. Earthquake size may be controlled by the location and size of fault patches, or by the amount of slip across these patches at the beginning of rupture. More research into the nature of rupture is required, however before a new hypothesis can be accepted.

Similar to earthquake magnitude, the \( \tau_{p}^{\text{max}} \) of an earthquake must be calculated from a waveform recorded by a seismometer. At any point between the fault plane and the seismometer, the \( \tau_{p}^{\text{max}} \) value may be altered. Site effects such as regional geology and structure may potentially be corrected for, but alterations to \( \tau_{p}^{\text{max}} \) caused by the nature of fault rupture cannot be removed. Taiwan is an ideal location to study possible site effects on \( \tau_{p}^{\text{max}} \) due to the large number of stations in the area, and frequent large earthquakes. Unfortunately not enough data from this region has been processed to allow an accurate calculation of station corrections. Evidence suggests site effects do influence the average \( \tau_{p}^{\text{max}} \) of an earthquake, but more work is required to understand the size of this influence.

While the previous two chapters represent significant improvements in our understanding of earthquake predominant period, there is still much more we need to learn. Now that \( \tau_{p}^{\text{max}} \) is observed as a characteristic of several earthquakes, researchers can study its
properties in a similar way to the study of magnitude, moment, rupture duration, and other earthquake characteristics.

Future research on $\tau_p^{\text{max}}$ from several different angles will hopefully lead to a more complete understanding of this phenomenon, and shed light on what causes the $\tau_p^{\text{max}}$ signal, and what mechanism controls earthquake size. Some potentially fruitful areas of research include: comparing fault type and rupture mechanism with $\tau_p^{\text{max}}$, studying $\tau_p^{\text{max}}$ in more world regions including Iran, Turkey, and Mexico, evaluating $\tau_p^{\text{max}}$ teleseismically, examining $\tau_p^{\text{max}}$ for small magnitude events, and fully evaluating instrument effects on $\tau_p^{\text{max}}$.

One of the first questions structural geologists ask when they hear about $\tau_p^{\text{max}}$ is if there is a relation between $\tau_p^{\text{max}}$ and fault type. Does an earthquake on a strike slip fault produce a $\tau_p^{\text{max}}$ value different from an event of the same magnitude on a thrust fault? A corollary to that question asks if events with different types of rupture, e.g. dip slip motion on a fault, versus strike slip motion on the same fault, produce different $\tau_p^{\text{max}}$ values. Answering both these questions require detailed study of well documented earthquakes, where the fault type and fault motion are both known. Whatever the answer, research in this direction will help develop a stronger understanding of $\tau_p^{\text{max}}$ behavior. Broadening the study regions of $\tau_p^{\text{max}}$ to a complete global dataset will help when the time comes to set up local early warning systems. Events from Taiwan, Japan, Alaska, and California all follow the same relationship between magnitude and period, but what about events from countries like Turkey, Iran, and Mexico? Recent large earthquakes devastated several regions of Iran, so installing an early
warning system based on $\tau_p^{\text{max}}$ could potentially save many lives. However, before such a system can be set up, the relationship between $\tau_p^{\text{max}}$ and magnitude must be verified in this region. Therefore pursuing this path of research will not only broaden the community’s understanding of $\tau_p^{\text{max}}$, but also potentially benefit the citizens of Iran.

One limitation currently imposed upon $\tau_p^{\text{max}}$ is that all stations used to calculate the value must be within 100km from the earthquake epicenter. Over longer distances, attenuation decreases the frequency content of the waveform, resulting in higher $\tau_p^{\text{max}}$ values which skew the predominant period trend. However, the maximum useful station distance is unknown, and study of attenuation’s effect on $\tau_p^{\text{max}}$ may allow stations of greater distance to be included in the calculation of $\tau_p^{\text{max}}$.

In addition to adding more stations at greater distances, a study of $\tau_p^{\text{max}}$ at teleseismic distances may find a different $\tau_p^{\text{max}}$ relationship. Perhaps there is a threshold frequency lost due to attenuation over a certain distance, so if all the $\tau_p^{\text{max}}$ measurements come from beyond that distance they will form a trend that relates to magnitude.

Numerous large magnitude events were added to the predominant period trend in the research presented in chapter 1. However no events with magnitude less than 5.9 were added to the trend. Additionally, all $\tau_p^{\text{max}}$ studies to date have only focused on events greater than magnitude 3.0. Part of the reason for this is the threshold noise levels, at which the noise is of similar amplitude to the event. A study of the predominant period for these small magnitude events may allow a lengthening of the magnitude period relationship below
magnitude 3.0, which would strengthen the overall relationship between predominant period and magnitude.

Earthquake prediction is a term to be used with great caution, but the possibility of finding a method for predicting earthquakes entices many geologists. $\tau_p^{\text{max}}$ is applicable to early warning, but may also offer a way to predict earthquakes. Unlike previous attempts to predict earthquakes, $\tau_p^{\text{max}}$ approaches the problem from a new front, that of frequency content, rather than amplitudes. The beginning of an earthquake is currently defined based on amplitudes. Once the amplitude of ground shaking exceeds the background noise level, an earthquake is said to begin. What if there was a change in frequency content before the change in amplitudes rises above the noise level? Then the earthquake would really begin prior to the amplitude trigger. If the frequency shift is significant, then the earthquake could be detected prior to the traditional trigger, providing fore-warning of ground shaking. This theory could be explored by examining the predominant period over a time window greater than the duration of the waveform. What is the frequency behavior immediately prior to the event? Is the frequency different than a background frequency recorded during quiescence? No one has examined this possibility yet, so the answer remains a mystery.
APPENDIX

Calculating the earthquake predominant period can be a useful step in analyzing the characteristics of an earthquake. $\tau_p^{\text{max}}$ of an earthquake contains information about the magnitude of an earthquake, along with clues about rupture properties. However, due to the nature of predominant period calculation, ensuring the correct input parameters is vital to achieving a good measurement. Some of these parameters can vary moderately without a drastic effect, while other parameters must be constrained to $1/100$th of a unit.

First, to compare $\tau_p^{\text{max}}$ from different events, uniformity is key. Unfortunately with a global data set, and several different magnitude earthquakes, a uniform data set is not easy to compile. Wave, the program that calculated $\tau_p^{\text{max}}$, is constructed to be as flexible as possible with the input data, but there are still limitations. All input data should be recorded at 100 samples/second, especially acceleration data. The need for this sampling rate comes from the real time conversions applied by Wave. In addition, all of the input data should come from stations within 100km of the event epicenter. This constraint is not one of Wave, but rather our current understanding of $\tau_p^{\text{max}}$ (see conclusions and future work).

Wave is designed to be a real-time program, able to calculate $\tau_p^{\text{max}}$ as an earthquake occurs. While this setup provides the most useful information for early warning, it results in a less than ideal setup for dealing with certain types of data. For acceleration data, rather than simply integrating the entire waveform, the conversion from acceleration to velocity must be done incrementally. This step by step conversion is performed by using an algorithm developed by Kanamori et al. [1999], and requires a few constants to be set prior to the execution of Wave. One such parameter is Q, which affects the filters applied in the
conversion from acceleration to velocity. Extensive testing of Q for 100 sample/second data
demonstrated a Q value of exactly 0.994 is best for performing the acceleration to velocity
conversion (Please refer to Kanamori et al. [1999] for a more detailed discussion of Q).

There is a trade off with an increase or decrease of Q which affects the similarity
between the predominant period calculated from broadband velocity and that calculated from
an acceleration derived velocity. This trade off was analyzed by comparing the predominant
period calculated from both a strong motion station, and a velocity sensor at the same
location, recording the same earthquake. At a Q of 0.994 the two predominant period traces
appear most similar, without artificially elevating the predominant period trace. If a
sampling rate different than 100 samples/second is to be used, then a new Q would have to be
determined by a similar trade off analysis.

In addition to Q, there are two other parameters set in Wave which can yield a
different $\tau_p^{\text{max}}$ result. The first parameter is the time delay before initiating the search for the
maximum point in the predominant period trace. The second parameter is the frequency of
the low pass filter applied to all data read into Wave. Here I will discuss the delay time, but
for a more detailed discussion of the frequency filter I refer the reader to Lockman and Allen
(in review).

The delay time in searching for $\tau_p^{\text{max}}$ was implemented to avoid measuring the
predominant period of the noise. Predominant period is calculated from the vertical velocity
trace using a smoothing constant (refer to the introduction for a more complete description).
This smoothing allows for a certain amount of information prior to the current time step to
affect the predominant period value at the current time step. This smearing allows the
possibility of the frequency of the noise to affect the frequency measured for the P-wave. 
After a small amount of time, the smearing is irrelevant, because the only data being 
combined is that from the P-wave. In the previous studies of $\tau_p^{\text{max}}$, researchers used a delay 
time of 0.5 seconds, or even 2.0 seconds in some cases [Allen and Kanamori, 2003; Lockman 
and Allen, In Review; Lockman and Allen, In Review]. However, more research into this 
delay time suggested that an accurate measurement of the predominant period is available in 
just 0.05 seconds after the trigger.

Shifting this delay time either earlier or later allows the user to focus on just one 
region of the predominant period trace. Many larger magnitude events reach their $\tau_p^{\text{max}}$ more 
than 1 second after the trigger, so moving the delay time earlier or later does not substantially 
affect the calculated $\tau_p^{\text{max}}$. However small magnitude events reach $\tau_p^{\text{max}}$ in a much shorter 
time, sometimes less than 0.1 seconds, so moving this delay time will effect the $\tau_p^{\text{max}}$ 
calculated for small magnitude events. Interestingly, moving the $\tau_p^{\text{max}}$ later, to 0.5 seconds, 
results in a steeper best-fit line, suggesting a more sensitive correlation between magnitude 
and period at this point (Figures 1 and 2). The reason for this has not yet been uncovered, 
but remains an area for potential future research.

References Cited

Allen, R.M., and H. Kanamori, The potential for earthquake early warning in southern 
Kanamori, H., P. Maechling, and E. Hauksson, Continuous monitoring of ground-motion 
Lockman, A., and R.M. Allen, Magnitude Period Scaling Relationships for Japan and the 
Pacific Northwest: Implications for Earthquake Early Warning, Bulletin of the 
Seismological Society of America, In Review.
Figure 1. $\tau_p^{\text{max}}$ versus magnitude plots for a range of delay times; a 0.00 seconds, b 0.05 seconds, c 0.10 seconds, d 0.50 seconds, e 1.00 seconds, and f 2.00 seconds, after the trigger. Note the change in the slope of the best fit line with increasing delay time.
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Figure 2. τ_d versus magnitude plots for a range of delay times; a 0.00 seconds, b 0.05 seconds, c 0.10 seconds, d 0.50 seconds, e 1.00 seconds, and f 2.00 seconds, after the trigger. With greater than 1.00 second delay, any meaningful trend is lost.