Proposed methodology for estimating the magnitude at which subduction megathrust ground motions and source dimensions exhibit a break in magnitude scaling: Example for 79 global subduction zones

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Abstract
In this article, I propose a method for estimating the magnitude ($M_B$) at which subduction megathrust earthquakes are expected to exhibit a break in magnitude scaling of both seismic source dimensions and earthquake ground motions. The methodology is demonstrated by applying it to 79 global subduction zones defined in the literature, including Cascadia. Breakpoint magnitude is estimated from seismogenic interface widths, empirical source scaling relations, and aspect ratios of physically unbounded earthquake ruptures and their uncertainties. The concept stems from the well-established observation that source-dimension and ground motion scaling decreases for shallow continental (primarily strike-slip) earthquakes when rupture exceeds the seismogenic width of the fault. Although a scaling break for megathrust earthquakes is difficult to observe empirically, all of the instrumentally recorded historical $M > 8.7$ mega-earthquakes have occurred on subduction zones with $M_B \approx 8.5$ ($8.1–8.9$), consistent with an observed break in source scaling relations derived from these same events. The breakpoint magnitudes derived in this study can be used to constrain the magnitude at which the scaling of ground motion is expected to decrease in subduction ground motion prediction equations.

Keywords
Megathrust earthquakes, subduction earthquakes, magnitude scaling, breakpoint magnitude, seismic hazard

Date received: 14 December 2019; accepted: 16 December 2019
Introduction

Because of the limited number of ground motion recordings that are available from subduction megathrust earthquakes, it is difficult to determine at what magnitude the magnitude scaling rate of radiated seismic ground motion (MSR) should begin to decrease. It also is not clear whether and to what extent this break in MSR should depend on the physical properties of the subduction zone. In this article, I propose a methodology for estimating breakpoint magnitudes that can be used to help constrain existing and future subduction ground motion prediction equations.

Background

Ground motions recorded during the 2010 moment magnitude (M) 8.9 Maule (Chile) earthquake and the 2011 M 9.0 Tohoku-Oki (Japan) earthquake are consistent with a break in MSR, where MSR becomes smaller at around M ≈ 7.6 – 8.2 according to the empirical ground motion models (GMMs) of Morikawa and Fujiwara (2013) and Abrahamson et al. (2016, 2018) and with the global data and GMM comparisons of Stewart et al. (2013). Other investigators have proposed that MSR decreases at smaller magnitudes (Ghofrani and Atkinson, 2014; Zhao et al., 2016; Zhao and Xu, 2012). However, all of these investigators agree that both the magnitude at which a scaling break occurs and the MSR beyond this break are not well constrained for megathrust earthquakes. Ghofrani and Atkinson (2014) show that their empirical GMM and that of Zhao et al. (2006), both based on Japanese recordings, have a similar mega-earthquake MSR to the global GMM of Abrahamson et al. (2016), which in turn is based on the finite-fault stochastic simulations of Gregor et al. (2002, 2006) and Atkinson and Macias (2009). The updated global GMM of Abrahamson et al. (2018) has a similar MSR to that of Abrahamson et al. (2016). Zhao and Xu (2012) empirically found a similar MSR to the finite-fault stochastic simulations of Gregor et al. (2002) for response-spectral accelerations recorded at short distances from mega-earthquakes in Japan.

The Japanese empirical GMM for subduction interface events developed by Morikawa and Fujiwara (2013), which incorporates ground motions from the 2011 Tohoku-Oki earthquake, required that MSR become independent of magnitude (i.e. MSR = 0) at M > 8.1 – 8.2 to explain the observed short-period ground motions from this and other M > 8 events in Japan. This is based largely on the similarity of ground motion amplitudes between the 2003 Tokachi-Oki (M 8.2) and Tohoku-Oki (M 9.0) earthquakes, although the authors admit that the reason why ground motion amplitudes should saturate at large magnitudes is unclear. Of course, very-long-period ground motions for these two earthquakes would not be expected to completely saturate based on the difference in their moment magnitudes. The empirical GMM of Zhao et al. (2016) for Japanese interface events supports a decrease in MSR at M > 7.1 to fit the Tohoku-Oki recordings but, as I show later, such a small MSR breakpoint magnitude is only controlled by the single Tohoku-Oki event and, therefore, is poorly constrained.

Kurahashi and Irikura (2011, 2013) found that five strong motion generation areas (SMGAs), delayed in time and spread out along strike (horizontal direction), are required to match the observed ground motions from the Tohoku-Oki earthquake with broadband simulations. This SMGA model is consistent with the amplitude saturation of short-period ground motions that were observed during the earthquake. These authors also observed that the amplitude and attenuation of recorded peak ground velocity (PGV) from the M 9.0 earthquake were virtually equivalent to that predicted from the GMM of Si and
Midorikawa (2000) for an earthquake of $M$ 8.0, whereas peak ground acceleration (PGA), although virtually equivalent to the empirical predictions at distances greater than 100 km, were larger than these predictions near the earthquake rupture. Wirth et al. (2017) and Frankel (2017) used the SMGA approach to successfully simulate broadband ground motions from the Tokachi-Oki and Maule earthquakes, respectively. Based on this success, Frankel et al. (2018) and Wirth et al. (2018) used this approach to simulate ground motions for a $M \approx 9$ earthquake on the Cascadia subduction zone (CSZ) not impacted by sedimentary basins in Seattle and Portland, which they showed are generally consistent with those predicted by the empirical GMM of Abrahamson et al. (2016).

It is clear from the above discussion that both empirical GMMs and stochastic and physics-based ground motion simulations predict that the MSR of subduction megathrust ground motions should decrease beyond some breakpoint magnitude. However, there is large uncertainty in this magnitude due to the paucity of mega-earthquakes with strong motion recordings. A better understanding of what causes such a break in MSR would help to determine how to constrain the magnitude at which it occurs for individual subduction zones. One approach is to look at the MSR of large shallow continental earthquakes for which there are a larger number of events above the magnitude at which a break in MSR is observed. Also, there are many empirical and theoretical studies that can be used to provide an empirical and physical basis for the observed break in MSR for such earthquakes.

In the discussion that follows, I review the empirical bases for a decrease in MSR at a specified breakpoint magnitude for shallow continental (predominantly strike-slip) earthquakes and follow this with a discussion of the relationship between this observed scaling break and a similar one observed in earthquake source scaling relations. This is followed by a discussion of how the same observation of a break in source scaling relations can be used to infer the magnitude at which a break in MSR is expected to occur for subduction megathrust earthquakes.

**Relationship between earthquake size and source properties**

Throughout the remainder of this article, the “size” of an earthquake is described either in terms of $M$ or seismic moment ($M_0$). Seismic moment is the preferred earthquake size metric of seismologists because of its relationship to the physical and dynamic properties of an earthquake. Seismic moment is defined by two different expressions depending on the data used to estimate it (e.g. Aki, 1972; Lay and Wallace, 1995):

$$M_0 = \mu AD = \mu LW \dot{D} \quad (1)$$

and

$$M_0 = 2\mu E_S / \Delta \sigma \quad (2)$$

where $\mu$ is the shear modulus (rigidity) of the crust, $A$ is the rupture area, $L$ is the rupture length, $W$ is the rupture width, $\dot{D}$ is the average rupture displacement or slip, $E_S$ is the radiated seismic energy, and $\Delta \sigma$ is the static stress drop. Equation 1 allows $M_0$ to be estimated from faulting parameters that can be observed in the field for large surface-rupturing earthquakes or derived from three-dimensional (3D) spatial aftershock
distributions. Equation 2 allows $M_0$ to be estimated from seismograms. Seismic moment has units of torque, usually given in N-m or dyn-cm (1 dyn-cm equals $10^7$ N-m).

Moment magnitude is defined in terms of $M_0$ (dyn-cm) by the following equation (Hanks and Kanamori, 1979):

$$M = \frac{2}{3} \log_{10} M_0 - 10.7$$

Therefore, the two earthquake size metrics are representations of the same physical properties of the source, which is why it has become the preferred earthquake magnitude metric used in GMMs worldwide.

**Fundamental properties of source scaling relations**

Because this article includes a discussion of source scaling relations, a brief summary of source scaling is warranted. Assuming a constant value of $D_s$ observed for a large range of seismic moments, Kanamori and Anderson (1975) used earthquake dislocation theory to show that $D \approx L$ and $M_0 \approx L^3 \times A^{3/2} \times T^3$ when rupture dimensions are unbounded by the seismogenic width of the fault ($W_{MAX}$), where $T$ is the source duration. The scaling with $T$ was not originally proposed by Kanamori and Anderson (1975) but, instead, comes from the definition of the far-field source spectrum proposed by Aki (1967) and Brune (1970). From Equation 3, this $M_0$ scaling corresponds to $M \propto \log_{10} L^2 \propto \log_{10} A \propto \log_{10} T^2$.

Scholz (1982a) proposes two models for earthquakes with large aspect ratios ($L > W_{MAX}$) in which rupture can only extend along strike. In the $W$-model, $\Delta \sigma \propto L$, $D \propto W_{MAX}$, $M_0 \propto L \propto A^{1/2} \propto T$, and $M \propto \log_{10} L^{2/3} \propto \log_{10} A^{1/3} \propto \log_{10} T^{2/3}$. In the $L$-model, $\Delta \sigma$ is constant as in the unbounded case, $D \propto L$, $M_0 \propto L^2 \propto A^2 \propto T^2$, and $M \propto \log_{10} L^{4/3} \propto \log_{10} A^{4/3} \propto \log_{10} T^{4/3}$.

Building on the work of Shaw and Scholz (2001), Shaw and Wesnousky (2008), and Manighetti et al. (2007), Shaw (2009) proposed a third source scaling model for large earthquakes that preserves the robust constant $\Delta \sigma$-scaling seen in small earthquakes:

$$D \propto \left(\frac{1}{L + \beta W_{MAX}}\right)^{-1}$$

$$M_0 \propto \frac{AL}{1 + \max(1, L/\beta W_{MAX})}$$

$$M \propto \log_{10} A + \frac{2}{3} \log_{10} \frac{L/A^{1/2}}{[1 + \max(1, L/\beta W_{MAX})]^{1/2}}$$

where $\beta$ is an empirical scaling parameter. These relations have the property of transitioning from an unbounded model to an $L$-model and finally to a $W$-model for $L > \beta W_{MAX}$ as the aspect ratio ($L/W_{MAX}$) increases.

**Large magnitude continental earthquakes**

In this section, I review empirical evidence for a break in MSR of large shallow continental earthquakes and relate it to a similar magnitude break in source scaling relations. This
is needed to set the premise for assuming that a similar correlation exists for megathrust earthquakes.

**MSR of continental earthquakes**

Zhao and Lu (2011) conducted an empirical study of the MSR of response-spectral acceleration from global shallow continental earthquakes. They observed a decrease in MSR at large magnitudes for these spectral amplitudes and proposed several physical and observational explanations for why this scaling break might occur, including the following: (1) the rupture length for large crustal earthquakes is usually very large, whereas the geometric and anelastic attenuation diminishes the amplitude of seismic waves generated from the distant part of the rupture at a recording station that is close to the fault; (2) the rupture propagation spreads the seismic waves from different parts of the rupture over a relatively long time period and increases the duration of the ground shaking at all recording stations; and (3) the amount of radiated energy increases with increasing magnitude up to $M \approx 7.0$ and then drops quickly for larger events.

In addition, comparisons of crustal GMMs from the NGA-West1 project (Abrahamson et al., 2008), the NGA-West2 project (Gregor et al., 2014), and the pan-European project (Campbell, 2016; Douglas et al., 2014) show a marked decrease in MSR at $M \approx 6.5 - 7.0$ for all spectral periods, although MSR is relatively steeper at longer periods. Some of these GMMs even predict near-zero MSRs at very close distances to the rupture surface for the larger events. This behavior is not new. It was empirically predicted for PGA at near-source distances four decades ago by Campbell (1981), who listed several empirical and theoretical studies conducted between 1969 and 1981 that supported the concept that PGA should become nearly independent of magnitude close to the rupture surface. This magnitude scaling behavior is consistent with the findings of Zhao and Lu (2011) as well as many theoretical studies, many of which offer physical explanations for this behavior (e.g. Baltay and Hanks, 2014; Douglas, 2002; Douglas and Jousset, 2011; Frankel, 2009; Halldorsson and Papageorgiou, 2005; Schmedes and Archuleta, 2008; Scholz, 1982a, 1982b; Somerville et al., 2006).

Some of the physical explanations that support an observed decrease in ground motion amplitudes for large magnitude earthquakes given by the above authors, some of which are similar to those of Zhao and Lu (2011), include the following: (1) a reduction in friction on the rupture surface once rupture propagation reaches steady state; (2) a decrease in dynamic stress drop for large magnitude earthquakes; (3) an inability of long ruptures to contribute to ground motions at sites located close to the rupture surface; (4) rupture of multiple asperities spread out in time and space during long rupture events; and (5) a reduction in source scaling between moment magnitude and rupture dimensions. Non-traditional support for a decrease in ground motion amplitudes at large magnitudes is a proposed breakdown in self-similarity for rupture displacements exceeding 1 m, as dynamic shear resistance becomes low, possibly due to friction-induced melting on the fault rupture surface (e.g. Di Toro et al., 2006).

**Source scaling relations of continental earthquakes**

One common explanation for a break in MSR is a break in source scaling. It is well established that small earthquakes scale according to the simple dislocation theory of Kanamori and Anderson (1975) when rupture dimensions are unbounded and $L/W \approx 1$ (e.g. Denolle
and Shearer, 2016; Gomberg et al., 2016; Hanks, 1977; Hanks and Bakun, 2002, 2008; Scholz, 1994; Wells and Coppersmith, 1994). However, there is considerable controversy how rupture dimensions of large continental earthquakes with $L/W > 1$ scale with earthquake size (e.g. Hanks and Bakun, 2014; Leonard, 2010; Manighetti et al., 2007; Shaw, 2013; Yin and Rogers, 1996, and references therein), in particular whether they exhibit scaling similar to the $L$-model, the $W$-model, the transition model of Shaw (2009), or something in between, such as the $M_0 \propto L^{5/2}$ scaling for $5.0 \leq M \leq 7.2$ events and $M_0 \propto L^{3/2}$ scaling for $M > 7.2$ events proposed by Leonard (2010, 2012, 2014).

A large number of seismologists (concentrating on the last few decades) have found or proposed that the change in source scaling at large magnitudes occurs as a result of rupture dimensions becoming bounded by the seismogenic width of the fault. This is based on a variety of global databases and source dimensions derived from aftershock distributions, seismological studies, and source inversions (Allen and Hayes, 2017; Anderson et al., 2017; Baltay and Hanks, 2014; Denolle and Shearer, 2016; Ellsworth, 2003; Gomberg et al., 2016; Hanks and Bakun, 2002, 2008, 2014; Konstantinou, 2014; Leonard, 2010, 2012, 2014; Mai and Beroza, 2000; Manighetti et al., 2007; Scholz, 1982a, 1982b, 1994; Shaw, 2009, 2013; Shaw and Scholz, 2001; Shaw and Wesnousky, 2008; Skarlatoudis et al., 2016; Stafford, 2014; Stirling et al., 2002; Stock and Smith, 2000; Yen and Ma, 2011). Ellsworth (2003) did not directly observe a scaling break in their rupture area data likely because they used only large strike-slip earthquakes; however, they found a similar scaling relation to that of Hanks and Bakun (2002, 2008) at large magnitudes, consistent with the $L$-model.

**Relationship between source and ground motion scaling**

Most of the observed breaks in source scaling for large continental strike-slip earthquakes referred to in the previous section occur within a magnitude range of $6.3 \leq M \leq 7.0$, consistent with a seismogenic crustal thickness of $W_{MAX} \approx 15 - 20$ km. This magnitude range is consistent with an abrupt change in MSR at $6.5 \leq M \leq 7.0$ observed in the modern empirical GMMs cited previously. The observed break in source scaling for large continental earthquakes is pervasive as is the MSR of modern GMMs. Stafford (2014) extended his probabilistic scaling model for strike-slip earthquakes to normal and reverse earthquakes and showed that the magnitudes at which saturation of rupture width occurs fall between 6.4–6.7 and 7.0–7.5, respectively. He suggested that the larger saturation magnitudes for reverse earthquakes are due to their relatively shallower dip angles and larger seismogenic widths than normal and strike-slip events.

The transition scaling relation of Shaw (2009) has been shown to provide a better fit to both simulated rupture displacements (Shaw and Scholz, 2001; Shaw and Wesnousky, 2008) and observed rupture displacements (Shaw, 2013) than either the $L$-model or the $W$-model. However, Shaw (2013) and Hanks and Bakun (2014) both demonstrated that there are only small differences in predicted rupture areas for continental strike-slip earthquakes between the $L$-model and the transition model for $M < 8.0$, above which there are very few continental events. Hanks and Bakun (2014) note that the transitional Shaw (2009) rupture area scaling relation, which is based on the same data as that of Hanks and Bakun (2008), transitions from an unbounded model to an $L$-model at $M \approx 6.3$ and then to a $W$-model at $M \approx 7.5$. According to Equation 4, this transition occurs because statistical regression of average slips found that $\beta = 6.9$ and $W_{MAX} = 15.6$ km, which forces the effective rupture width to be 6.9 times larger than $W_{MAX}$, or about 108 km, before $W$-model scaling occurs. Shaw and Wesnousky (2008) and Shaw (2009) suggest that this larger effective width is due
Scholz (1982b), using simple dislocation theory, demonstrated that the $W$-model implies that PGA and PGV scale linearly with $L$, whereas the $L$-model implies that PGA scales as $\sqrt{\log_{10}L}$ and PGV scales as $\sqrt{L}$. This latter scaling model predicts a smaller MSR and smaller ground motion amplitudes at large magnitudes than the $W$-model and is more consistent with MSRs of modern GMMs (see previous citations). It also demonstrates that a break in source scaling can impact the amplitude of strong ground motion.

**Large magnitude subduction earthquakes**

This brings me back to a discussion of the MSR of subduction megathrust earthquakes and its hypothesized relationship to source-dimension scaling. First, I review observed and proposed source scaling relations for subduction megathrust earthquakes. This is followed by a proposal to use these scaling relations to estimate the magnitudes at which a break in MSR is expected to occur for megathrust events.

**Source scaling relations of megathrust earthquakes**

A search of the literature discovered 11 publications in the last two decades that can be used to estimate source dimensions from $M_0$ or $M$ for megathrust earthquakes: Allen and Hayes (2017) (hereafter AH17), Blaser et al. (2010) (hereafter Bea10), Brengman et al. (2019) (hereafter Bea19), Goda et al. (2016) (hereafter Gea16), Leonard (2014) (hereafter L14), Murotani et al. (2013) (hereafter Mea13), Papazachos et al. (2004) (hereafter Pea04), Skarlatoudis et al. (2016) (hereafter Sea16), Strasser et al. (2010) (hereafter Sea10), Thingbaijam et al. (2017) (hereafter Tea17), and Ye et al. (2016) (hereafter Yea16). Sea16 supersedes an earlier study by Somerville et al. (2002), L14 supersedes an earlier study by Leonard (2010, 2012), and Mea13 supersedes an earlier study by Murotani et al. (2008). It is beyond the scope of this study to review each of these scaling relations. A close review of the largest megathrust earthquakes in these studies indicates there is a possible scaling break at $M \approx 8.5$ in one or more of the source dimensions. The strongest indication of a magnitude scaling break is for $W$ because of its tendency to totally saturate once the fault width has been exceeded, which is why Sea16 and AH17 proposed a bilinear scaling relation for rupture width and AH17 with rupture area. This indication is weaker for $L$ and $A$ because the relations for these parameters only change slope beyond the breakpoint. Nevertheless, there are only a few events above this break which is why many authors are reluctant to propose one. In addition, some of the studies used source dimensions estimated by different authors and methodologies that might mask a scaling break if one exists.

There are a few studies of megathrust earthquakes that, although they did not propose source scaling relations, are relevant to the behavior of such earthquakes. Denolle and Shearer (2016) estimated the $P$-wave displacement source spectra of 942 shallow global continental reverse and subduction megathrust earthquakes of $5.5 \leq M \leq 9.2$. One of their major findings was that the shapes of the source spectra for the larger earthquakes were consistent with a double corner frequency source spectrum with an intermediate frequency falloff of $f^{-1}$ and a high-frequency falloff of $f^{-2}$. The first corner frequency relates closely to the source duration $T$. This change in shape occurs at $M \approx 7.5$ and is consistent with a
transition from $M_0 \propto T^3$ (unbounded constant stress drop scaling) to $M_0 \propto T^2$ (L-model scaling). The authors suggested that this observed break in scaling is consistent with an increase in rupture aspect ratio likely caused when rupture saturates the full seismogenic width of the fault and continues to rupture along strike. It is similar to a break in scaling observed for continental reverse earthquakes by Stafford (2014), which suggests that this break is not related to the megathrust events in their database but to the continental reverse earthquakes. Their plot of estimated source dimensions suggests that $W$ saturates above $M \approx 8.5$, which is consistent with a similar scaling break in other megathrust scaling relations. They also suggest that $\Delta \sigma$ and scaled energy ($E_S/M_0$) are relatively constant over all magnitudes implying self-similarity even though their source geometry and spectral shapes vary with earthquake size.

Corbi et al. (2017) observed that the largest megathrust earthquakes share a common characteristic: they rupture more than one asperity along their strike. To investigate the role of asperity sizes and spacing on maximum magnitude and percentage of synchronized ruptures, these authors used analog models to simulate along-strike rupture behavior of megathrust earthquakes. They found that the barrier-to-asperity length ratio had a negative correlation with maximum event magnitude. A barrier is an area of the fault where rupture is inhibited, whereas an asperity is an area where, once initiated, rupture is free to propagate due to velocity-weakening frictional behavior. They observed a permanent barrier behavior for a barrier-to-asperity length ratio greater than 0.5, whereas an earthquake can rupture one or more barriers resulting in a larger magnitude earthquake for smaller barrier-to-asperity length ratios. The rupture of multiple barriers is likely to be associated with ruptures with larger aspect ratios and is consistent with the concept of a “runaway” earthquake proposed by Woo and Mignan (2018).

Meier et al. (2017) used a catalog of source time functions to model the typical temporal rupture behavior of large ($M > 7.0$) megathrust earthquakes. They observed linear and magnitude-independent seismic moment rate that implies source duration scales as $M_0 \propto T^2$, consistent with the L-model, although the data range is not sufficient to confirm or rule out $M_0 \propto T^3$ scaling, which they note is well established for smaller unbounded megathrust earthquakes. Nevertheless, these authors proposed that there is a magnitude threshold at which the source duration scaling changes with earthquake size. They note that this transition must occur before the linear trend in source time functions is well constrained, which they suggest occurs at source dimensions smaller than the seismogenic width of typical subduction zones. They suggested that this could occur due to plausible weakening mechanisms, such as fault melting or thermal pressurization, which require a minimum energy level to be triggered. Their plots of inverse initial slope, characteristic time scale, and centroid time of the source time functions suggested a scaling break at $M \approx 8.5$ similar to that found in source dimensions of other studies of megathrust earthquakes.

**Inference for a break in MSR of subduction ground motions**

I propose that the observed break in source scaling relations for subduction megaeathquakes at $M \approx 8.5$ presented in the previous section can be used to infer a similar break in MSR, as demonstrated for continental strike-slip and, to a lesser extent, reverse earthquakes. However, before doing that, I would like to address some apparent inconsistencies in subduction scaling relations that came up in the previous section.

The discrepancy in $D$ scaling noted by Denolle and Shearer (2016) for megaeathquakes might be explained by the transition scaling relation of Shaw and Scholz
These authors used a 3D dynamic earthquake model to simulate how $\bar{D}$ should scale with aspect ratio. What they found was that instead of $\bar{D}$ becoming constant when $W$ is bounded by the width of the seismogenic zone ($W$-model), leading to an increase in aspect ratio first proposed by Kanamori and Anderson (1975), there is a wide range of aspect ratios over which the transition from $\bar{D} \propto L$ scaling to constant $\bar{D}$ scaling occurs. Their simulations showed that this transition occurs between aspect ratios of about 2–10 and is relatively consistent with the scaling relations proposed by Shaw (2009, 2013) for continental strike-slip earthquakes that predict this transition occurs at an aspect ratio of about 7. Yin and Rogers (1996) suggested that this transition happens at an aspect ratio of about 5 and Stock and Smith (2000) at an aspect ratio of about 4. All of these studies indicate that it takes quite a long distance and range of magnitudes for $\bar{D}$ to reach its maximum value during dynamic rupture, which might explain why $L$-model scaling rather than the expected $W$-model scaling is observed when rupture width saturates. For example, an aspect ratio of 7 and an average seismogenic interface width of 100 km are consistent with a transition from $L$-model to $W$-model scaling at $L \approx 700$ km, longer than most historical megathrust earthquakes.

I would expect the larger barrier-to-asperity length ratios that Corbi et al. (2017) associates with larger megathrust earthquakes to be encountered once an event has ruptured or nearly ruptured the full width of the subduction interface and continues to rupture along strike, similar to that observed for fault-width bounded continental earthquakes. Once that occurs, the rupture is expected to transition to $L$-model and then to $W$-model scaling which, like continental earthquakes, is expected to be accompanied by a decrease in MSR. Even if $\bar{D}$ does not exactly become constant at $L/W > 1$ ($W$-model) as found by Shaw and Scholz (2001) and Shaw (2009, 2013), the fact that the rate of scaling of $\bar{D}$ with seismic moment decreases with increasing aspect ratio suggests that ground motions should decrease accordingly, resulting in a corresponding decrease in MSR.

Ji and Archuleta (2018) clearly demonstrated that rupture dimensions of subduction intraslab events constrained by the width of the oceanic slab will cause a break in MSR. They used kinematic broadband simulations to estimate the values of PGA and PGV to distances of 256 km for $M_{\text{6.2}}, M_{\text{6.8}}, M_{\text{7.4}},$ and $M_{\text{8.0}}$ intraslab events occurring within the subducting Japan and Cascadia oceanic slabs. The scenarios were assumed to have a normal fault mechanism and an average stress drop of 10 MPa, consistent with most intraslab earthquakes. Hypocenter locations, slip distributions, rupture dip angles, source-to-site azimuths, and slab thicknesses were varied to account for epistemic uncertainty in the ground motion estimates. Their simulations showed that the thicker Japan slab (36–76 km) indicates a possible decrease in MSR of PGA and PGV between the $M_{\text{7.4}}$ and $M_{\text{8.0}}$ scenario earthquakes, consistent with the assigned breakpoint (which the authors call saturation) magnitudes that range from 7.6 to 8.2. The thinner Cascadia slab (9–21 km) shows a decrease in MSR for PGA and PGV between the $M_{\text{6.8}}$ and $M_{\text{7.4}}$ earthquake scenarios, consistent with the assigned range in breakpoint magnitudes of 6.3–7.1. They attributed these differences to the maximum brittle (seismogenic) rupture width allowed by the two subducting slabs and went on to use these results to estimate the expected breakpoint magnitudes for several global subduction zones based on oceanic slab thicknesses available in the literature.

I propose that the observation by many investigators that there is a break in source scaling for continental strike-slip earthquakes at $M_{\text{6.5}},$ continental reverse earthquakes at $M_{\text{7.5}},$ and megathrust earthquakes at $M_{\text{8.5}},$ and that this break is likely due to saturation of the seismogenic width of the fault, can be used to extrapolate the results of Ji
and Archuleta (2018) for intraslab earthquakes to interface earthquakes. The estimated seismogenic width of the subduction interface is used to estimate this breakpoint magnitude. The assumption is that this breakpoint magnitude should also correspond to the magnitude associated with a break in MSR of megathrust earthquakes similar to that found for continental strike-slip and reverse events. An estimate of this breakpoint magnitude for major subduction zones around the world is presented in the next section.

Although there are other physical explanations that might explain a decrease in MSR proposed for continental and megathrust earthquakes, I invoke Ockham’s razor which, as applied to the field of mathematics by Jefferys and Berger (1991), dictates that the simplest plausible explanation should be used to explain a given phenomenon. I suggest that the simplest explanation for the observed break in source scaling and MSR is when the rupture saturates the width of the subduction interface and continues to rupture along strike. Using the seismogenic width of the subduction interface to represent when this break occurs is considered an upper-bound estimate of breakpoint magnitude because of other complexities in the rupture process that might cause some events to transition to one-dimensional (1D) scaling before their rupture fills the entire width of the interface (e.g., Corbi et al., 2017). This is taken into account to some extent by using a range of aspect ratios in the calculations.

**Breakpoint magnitude for megathrust earthquakes**

Building on the hypothesis formulated in the previous section, three sets of data were used to estimate the magnitude at which a proposed break in source scaling and MSR is expected to occur for megathrust earthquakes: (1) estimates of the seismogenic width of the subduction interface; (2) scaling relations between earthquake size and rupture dimensions for megathrust earthquakes; and (3) aspect ratios of megathrust events unconstrained by the width of the subduction interface. Epistemic uncertainties in these data are used to construct a logic tree to provide a distribution of breakpoint magnitudes for each subduction zone. The logic tree is divided into four major branches: interface width, source scaling model (publication), source scaling relations for each model, and aspect ratio, along with their associated weights. When a normal (Gaussian) distribution is assumed, weights of 0.185, 0.630, and 0.185 are used to define a three-point normal distribution represented by its 5% confidence limit (CL), its mean and median (50th percentile), and its 95% CL, respectively (Keefer and Bodily, 1983).

**Width of subduction interface**

Berryman et al. (2015) compiled a database of global subduction zone interface properties as part of the *Global Earthquake Model (GEM) Faulted Earthquake Subduction Interface Characterisation Project*. A group of renowned international seismologists used available data on subduction zone properties together with analysis and expert opinion to provide, among other properties, the estimated minimum, preferred, and maximum seismogenic interface widths of 79 global subduction zones and their major segments. I used the preferred value for each subduction zone because the minimum and maximum estimates calculated by Berryman et al. (2015) by combining the minimum and maximum estimates of fault dip with the minimum and maximum depths to the top and bottom of the seismogenic interface produced extreme estimates of interface width in many cases. I refer to the preferred seismogenic width as $W_{MAX}$ in the remainder of this article. As shown later, the largest 95% CL breakpoint magnitude estimated for any subduction zone in this study is
M 9.7, very close to the maximum magnitude of M 9.6 estimated by Berryman et al. (2015) for all 79 subduction zones. Hayes et al. (2012) and Hayes (2018) also provide global subduction zone properties, but the compilation of Berryman et al. (2015) is the most complete and easiest to apply.

Due to the critical nature of seismogenic width in estimating breakpoint magnitude using the proposed methodology, it is important to know how the preferred widths were derived. Berryman et al. (2015) state that their preferred seismogenic widths were calculated using preferred estimates of the dip angle, up-dip depth, and down-dip depth of the seismogenic subduction interface. For subduction zones where special studies had been conducted using seismicity, geodetic, geophysical, and/or thermal data, these data were used to estimate these parameters. Otherwise, they were determined from the Hayes et al. (2012) database of global subduction zone geometries, which were derived primarily from the locations and focal mechanisms of good quality interface thrust events. Where the authors had no information about the dip angle, they used a default value of 15°. In all cases, the authors used the intersection of the trench with the Earth’s surface as a default minimum up-dip depth to account for the possibility that rupture to the trench cannot be ruled-out anywhere. They used 5 km below the intersection of the subduction interface and the seafloor as a default preferred value when no other information was available. For subduction zones where the authors did not have knowledge of down-dip depth, they assumed a default value of 35 km. Because of a general paucity of interface earthquakes, the CSZ was one subduction zone for which Berryman et al. (2015) used a regional study as described in a subsequent section of this article.

Source dimensions

Relationships between earthquake size (M₀ or M) and rupture dimensions of megathrust earthquakes are required to estimate earthquake size from interface width. I use W-scaling relations to directly estimate earthquake size from \( \bar{W}_{MAX} \). In addition, I use A- and L-scaling relations to capture additional epistemic uncertainty by increasing the number of scaling relations that can be used to estimate M₀ and M. In this latter case, estimates of aspect ratio are required to calculate M from values of A and L that are consistent with \( \bar{W}_{MAX} \), which I refer to as \( A_{MAX} \) and \( L_{MAX} \). These aspect ratios are discussed in the next section.

The 11 source scaling models presented previously are used to estimate source parameters. Of these, 8 provide scaling relations for W (Pea04, Bea10, L14, Gae16, Sea16, AH17, Tea17, Bea19), 10 provide scaling relations for A (Pea04, Sea10, Mea13, L14, Gae16, Sea16, Yea16, AH17, Tea17, Bea19), and 8 provide scaling relations for L (Pea04, Bea10, Sea10, L14, Gae16, AH17, Tea17, Bea19). In the logic tree, equal weights were assigned to each scaling model and to each scaling relation within each model. The equal weight assigned to each scaling model assumes that each study is equally valid. The equal weight assigned to each scaling relation recognizes the fact that each study uses a common database and methodology to develop the scaling relations. Moment magnitude is estimated using Equation 3 for those relations that define earthquake size in terms of seismic moment.

Sea16 and AH17 proposed alternative linear and bilinear models for rupture width, but only the bilinear relation is used for AH17 as recommended by these authors. The bilinear models predict that W saturates with magnitude at M > 8.4 (Sea16) and M > 8.7 (AH17), based on the estimated rupture widths of the mega-earthquakes included in their databases. They proposed that this saturation might be an indication of fault-width saturation,
which they suggested most likely occurs at $W_{\text{MAX}} \approx 200 \text{ km}$, consistent with the proposal of Tajima et al. (2013). They also state that this width might vary from one subduction zone to another. AH17 also proposed alternative bilinear and linear models for $A$ for $M > 8.6$, but only the bilinear relation was used as recommended by AH17. Bea10 proposed alternative models for $W$ and $L$ using only oceanic (subduction interface) events and combined continental reverse and oceanic interface events. L14 and Yea16 included both continental reverse and subduction interface events in their models. Because only two scaling models provided alternative bilinear (i.e. width saturation) $W$-scaling relations, I checked whether these alternative relations resulted in a significant difference in estimated values of breakpoint magnitude as a test of whether estimates of $M_B$ might be biased by the linear scaling assumption assumed by most of the seismologists. The differences were found to be less than 0.1 magnitude units because of the shallower slope between log ($W$) and $M$ in the linear relation that mitigated the differences in $M_B$ at the larger magnitudes of interest in this study.

### Aspect ratios

Typical aspect ratios of earthquakes that approximately rupture the full seismogenic width of the subduction interface without continuing to rupture along strike are required to estimate earthquake size from $A$ and $L$. I used the two bilinear $W$-scaling relations (Sea16 and AH17) referenced in the previous section to estimate these aspect ratios. This avoids potentially overestimating the aspect ratios with unreasonably large rupture widths. To accomplish this, I directly estimated $M_{\text{MAX}}$ from the minimum of $W_{\text{MAX}}$ or the saturation width in these relations for each of the 79 subduction zones. I then estimated $L(M_{\text{MAX}})$ (AH17 only) and $A(M_{\text{MAX}})$ (both Sea16 and AH17) from this estimate of $M_{\text{MAX}}$ using the $L$- and $A$-scaling relations. Aspect ratios were then estimated as either $AR_{\text{MAX}} = L(M_{\text{MAX}})/W_{\text{MAX}}$ or $AR_{\text{MAX}} = A(M_{\text{MAX}})/W_{\text{MAX}}^2$. The $W$-scaling relation of Sea16 had to be inverted to predict $M_{\text{MAX}}$ from $W_{\text{MAX}}$. Using inverted relations when they have not been developed using orthogonal regression is not statistically robust, but I considered it necessary to capture epistemic uncertainty in the results. AH17 used orthogonal regression mitigating the need to invert their equations. This analysis gave a mean aspect ratio of 1.52 with a range of 0.99–2.04. Using this result, logic-tree values of the 5% CL, mean, and 95% CL aspect ratios ($AR_{\text{MAX}}$) were defined as 1.0, 1.5, and 2.0, respectively, with weights of 0.185, 0.630, and 0.185. Rupture length and rupture area estimates consistent with $W_{\text{MAX}}$ were then calculated as $A_{\text{MAX}} = AR_{\text{MAX}}^2 W_{\text{MAX}}^2$ and $L_{\text{MAX}} = A_{\text{MAX}} W_{\text{MAX}}$ and assigned the same weights as assigned to $AR_{\text{MAX}}$.

### Breakpoint magnitudes

I used values of $W_{\text{MAX}}, A_{\text{MAX}},$ and $L_{\text{MAX}}$ together with the source scaling models and scaling relations and the weights presented previously to estimate values of $M_B$ for all 79 subduction zones. An evaluation of this logic tree resulted in a total of 68 estimates of $M_B$ for each subduction zone. The values of $W_{\text{MAX}}$ along with the weighted mean ($M_B$), weighted median ($M_B$), weighted epistemic standard deviation ($\sigma_B$), weighted 5% CL ($M_B^{\text{5\%}}$), and weighted 95% CL ($M_B^{\text{95\%}}$) of the 68 breakpoint magnitudes for each subduction zone are listed in Table 1. Average statistics for all of the subduction zones are given in Table 2. In the remainder of this article, the plus and minus epistemic standard deviation values are displayed as the mean “±” standard deviation (e.g. 8.0±0.25), the 90% confidence bounds are displayed as the mean with a subscript (5% CL – mean) and superscript (95% CL –
Table 1. Estimated breakpoint magnitudes of global subduction zones defined by Berryman et al. (2015).

<table>
<thead>
<tr>
<th>Subduction zone</th>
<th>$W_{MAX}$</th>
<th>$M_B$</th>
<th>$\sigma_B$</th>
<th>$M_B$</th>
<th>$M_B^{5%}$</th>
<th>$M_B^{95%}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Alaska-Aleutians</td>
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<td>0.26</td>
<td>8.2</td>
<td>7.9</td>
<td>8.8</td>
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<td>0.25</td>
<td>8.0</td>
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<td>8.5</td>
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<tr>
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<td>8.8</td>
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<td>8.4</td>
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<td>0.26</td>
<td>8.3</td>
<td>8.0</td>
<td>8.9</td>
</tr>
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<td>9.0</td>
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<td>8.1</td>
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<tr>
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<td>7.6</td>
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<td>8.0</td>
<td>7.7</td>
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<td>8.6</td>
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<tr>
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<td>8.6</td>
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<tr>
<td>East</td>
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<td>8.0</td>
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<td>8.6</td>
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<tr>
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<td>8.6</td>
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<td>Andaman-Sunda Trench</td>
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<tr>
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<td>Cyprus (Eastern Segment)</td>
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</tbody>
</table>

(continued)
mean) (e.g. 8.0$^{+0.6}_{-0.5}$), and the range of minimum and maximum values are displayed in parentheses following the mean (e.g. 8.0 (7.5–8.5)).

**Discussion**

Table 2 shows that there is a wide range of breakpoint magnitudes for the 79 subduction zones defined by Berryman et al. (2015). It also shows that there is a relatively large epistemic standard deviation and 90% confidence bounds. Unfortunately, it is difficult to independently validate the values of $M_B$ listed in Table 1 because no similar study exists. However, they are shown to be consistent with other complementary information and data available in the literature in the sections that follow.

**Comparison with historical mega-earthquakes**

One important consistency check is to compare the magnitudes and rupture widths of the largest instrumentally recorded mega-earthquakes with the estimated breakpoint magnitudes and seismogenic interface widths of the subduction zones on which they occurred. For this purpose, I used the magnitudes and rupture dimensions of the $M \geq 8.7$ mega-earthquakes included in 8 of the 11 source scaling models adopted for this study, for which

<table>
<thead>
<tr>
<th>Subduction zone</th>
<th>$W_{MAX}$</th>
<th>$M_B$</th>
<th>$\sigma_B$</th>
<th>$\bar{M}_B$</th>
<th>$M_{5%}^B$</th>
<th>$M_{95%}^B$</th>
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</thead>
<tbody>
<tr>
<td>South America</td>
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<td>0.28</td>
<td>8.5</td>
<td>8.2</td>
<td>9.2</td>
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<td>9.2</td>
</tr>
<tr>
<td>Peru</td>
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<td>8.6</td>
<td>8.2</td>
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<tr>
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<td>0.25</td>
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<td>0.29</td>
<td>7.3</td>
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</tr>
<tr>
<td>El Salvador-Nicaragua</td>
<td>42</td>
<td>7.3</td>
<td>0.29</td>
<td>7.3</td>
<td>6.8</td>
<td>7.7</td>
</tr>
<tr>
<td>Costa Rica-West Panama</td>
<td>68</td>
<td>7.7</td>
<td>0.25</td>
<td>7.7</td>
<td>7.3</td>
<td>8.2</td>
</tr>
<tr>
<td>Antilles</td>
<td>99</td>
<td>8.1</td>
<td>0.25</td>
<td>8.0</td>
<td>7.7</td>
<td>8.6</td>
</tr>
<tr>
<td>Manila</td>
<td>100</td>
<td>8.1</td>
<td>0.25</td>
<td>8.0</td>
<td>7.7</td>
<td>8.6</td>
</tr>
<tr>
<td>Philippine</td>
<td>57</td>
<td>7.6</td>
<td>0.26</td>
<td>7.6</td>
<td>7.2</td>
<td>8.0</td>
</tr>
<tr>
<td>East Luzon Trough</td>
<td>73</td>
<td>7.8</td>
<td>0.25</td>
<td>7.7</td>
<td>7.3</td>
<td>8.3</td>
</tr>
<tr>
<td>Cotabato Trench</td>
<td>100</td>
<td>8.1</td>
<td>0.25</td>
<td>8.0</td>
<td>7.7</td>
<td>8.6</td>
</tr>
<tr>
<td>Sulu Trench</td>
<td>100</td>
<td>8.1</td>
<td>0.25</td>
<td>8.0</td>
<td>7.7</td>
<td>8.6</td>
</tr>
<tr>
<td>Minahassa Trench</td>
<td>102</td>
<td>8.1</td>
<td>0.25</td>
<td>8.0</td>
<td>7.7</td>
<td>8.6</td>
</tr>
<tr>
<td>Seram Trough</td>
<td>93</td>
<td>8.0</td>
<td>0.25</td>
<td>8.0</td>
<td>7.6</td>
<td>8.5</td>
</tr>
<tr>
<td>Timor</td>
<td>104</td>
<td>8.1</td>
<td>0.25</td>
<td>8.1</td>
<td>7.7</td>
<td>8.6</td>
</tr>
<tr>
<td>Manokwari Trench</td>
<td>100</td>
<td>8.1</td>
<td>0.25</td>
<td>8.0</td>
<td>7.7</td>
<td>8.6</td>
</tr>
<tr>
<td>Halmahera</td>
<td>108</td>
<td>8.1</td>
<td>0.25</td>
<td>8.1</td>
<td>7.8</td>
<td>8.6</td>
</tr>
<tr>
<td>Kepulauan Sangihe</td>
<td>108</td>
<td>8.1</td>
<td>0.25</td>
<td>8.1</td>
<td>7.8</td>
<td>8.6</td>
</tr>
</tbody>
</table>

$W_{MAX}$ is the preferred seismogenic interface width (km); $M_B$, $\bar{M}_B$, $\sigma_B$, $M_{5\%}^B$, and $M_{95\%}^B$ are the epistemic mean, median, standard deviation, 5% CL, and 95% CL of 68 estimated breakpoint magnitudes per subduction zone segment, respectively.
data were available in an electronic supplement or in databases in the publication or that the authors were kind enough to provide. Two additional models included magnitudes but not rupture widths. Not all of the models included data for all of the mega-earthquakes. I included the 1700 Cascadia earthquake based on the comprehensive study of its source properties by Satake et al. (2003) described below and in subsequent sections of this article.

If my hypothesis is true, I should find that the earthquake magnitudes are larger than $M_B$, that their rupture widths are generally consistent with $W_{MAX}$, and that their aspect ratios are larger than unity. For this purpose, I included the minimum and maximum values of seismogenic interface widths from Berryman et al. (2015) to define the range associated with $W_{MAX}$. This comparison is shown in Table 3.

The 1700 Cascadia and 2005 Nias-Simeulue earthquakes included in Table 3 warrant further discussion. The 2005 Nias-Simeulue earthquake originally appeared to be an outlier given that its mean (218 km) and range (201–260 km) of rupture widths fall completely outside of the corresponding range of $W_{MAX}$ estimates. All of the estimates of $W$ were derived

<table>
<thead>
<tr>
<th>Earthquake</th>
<th>Year</th>
<th>Subduction zone</th>
<th>Historical earthquake data</th>
<th>Subduction data</th>
</tr>
</thead>
<tbody>
<tr>
<td>Cascadia</td>
<td>1700</td>
<td>Cascadia</td>
<td>9.0 (8.8–9.2) 74 (60–90) 15 (12–18)</td>
<td>68 (30–106) 7.7 (7.3–8.2)</td>
</tr>
<tr>
<td>Kamchatka</td>
<td>1952</td>
<td>Japan-Kuril:</td>
<td>8.9 (7.8–9.0) 168 (130–210) 3 (2–4)</td>
<td>134 (87–189) 8.3 (8.0–8.9)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Kuril-Kamchatka</td>
<td>(8.7–9.0) (160–175) (3–4)</td>
<td></td>
</tr>
<tr>
<td>Andrewof Islands</td>
<td>1957</td>
<td>Alaska-Aleutians: Western Aleutians</td>
<td>8.6 (8.4–8.7) 106 (90–128) 7 (6–9)</td>
<td>123 (86–177) 8.3 (7.9–8.8)</td>
</tr>
<tr>
<td>Valparaiso</td>
<td>1960</td>
<td>South America:</td>
<td>9.3 (8.8–9.9) 192 (157–240) 5 (3–6)</td>
<td>192 (130–265) 8.7 (8.3–9.3)</td>
</tr>
<tr>
<td>Rat Islands</td>
<td>1965</td>
<td>Alaska-Aleutians: Western Aleutians</td>
<td>8.7 (8.7) 108 (100–115) 5 (5–6)</td>
<td>123 (85–185) 8.3 (7.9–8.8)</td>
</tr>
<tr>
<td>Prince William Snd.</td>
<td>1964</td>
<td>Alaska-Aleutians: Prince William Snd.</td>
<td>9.2 (9.1–9.3) 226 (180–266) 3 (3–4)</td>
<td>263 (200–300) 8.9 (8.4–9.7)</td>
</tr>
<tr>
<td>Sumatra-Andaman</td>
<td>2004</td>
<td>Andaman-Sunda:</td>
<td>9.1 (9.0–9.2) 189 (175–201) 6 (3–8)</td>
<td>145 (99–174) 8.4 (8.0–9.0)</td>
</tr>
<tr>
<td>Nias-Simeulue</td>
<td>2005</td>
<td>Andaman-Sunda:</td>
<td>8.6 (8.4–8.8) 110 (90–130) 3 (2–3)</td>
<td>103 (50–186) 8.1 (7.7–8.6)</td>
</tr>
<tr>
<td>Maule</td>
<td>2010</td>
<td>South America:</td>
<td>8.9 (8.7–9.0) 193 (179–249) 3 (2–3)</td>
<td>192 (130–265) 8.7 (8.3–9.3)</td>
</tr>
<tr>
<td>Tohoku-Oki</td>
<td>2011</td>
<td>Japan-Kuril:</td>
<td>9.1 (9.0–9.1) 189 (176–203) 2 (2–3)</td>
<td>166 (128–224) 8.5 (8.1–9.2)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Japan</td>
<td>(8.7–9.0) 189 (176–203) 2 (2–3)</td>
<td></td>
</tr>
</tbody>
</table>

$^a$Mean (range) of event magnitude, $M$, rupture width (km), $W$, and rupture aspect ratio, $AR$.

$^b$Mean (5% to 95% CL) of seismogenic interface width (km), $W_{MAX}$, and breakpoint magnitude, $M_B$. 

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Table 2. Global summary statistics of breakpoint magnitudes for subduction zones listed in Table 1.

<table>
<thead>
<tr>
<th>Statistic</th>
<th>$M_b$</th>
<th>$\sigma_b$</th>
<th>$\bar{M}_b$</th>
<th>$\bar{M}_b$</th>
<th>$\bar{M}_b$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Minimum</td>
<td>7.0</td>
<td>0.25</td>
<td>7.1</td>
<td>6.3</td>
<td>7.4</td>
</tr>
<tr>
<td>Maximum</td>
<td>8.9</td>
<td>0.35</td>
<td>8.9</td>
<td>8.4</td>
<td>9.7</td>
</tr>
<tr>
<td>Mean</td>
<td>7.9</td>
<td>0.26</td>
<td>7.9</td>
<td>7.5</td>
<td>8.4</td>
</tr>
<tr>
<td>Median</td>
<td>8.0</td>
<td>0.25</td>
<td>8.0</td>
<td>7.6</td>
<td>8.5</td>
</tr>
</tbody>
</table>

See footnote to Table 1 for definition of parameters.

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Table 3. Comparison of historical $M$=8.7 subduction interface mega-earthquake magnitudes and rupture widths with breakpoint magnitudes and seismogenic interface widths proposed in this study.

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Campbell
from finite-fault inversions available in the Finite-Source Rupture Model Database SRCMOD (Mai and Thingbaijam, 2014). Upon further research, I discovered that these inversion results are not consistent with the locations of well-constrained shallow thrust aftershocks that occurred along the subduction interface (Tilmann et al., 2010) or with the seismogenic width of the Sumatra megathrust in the vicinity of the earthquake determined from its geophysical characteristics (Meltzner et al., 2012; Shulgin et al., 2013). The well-constrained shallow thrust aftershocks define a rupture width of about 110 km with a range of 90–130 km. These latter values are included in Table 3 and were used as the preferred widths in the hypothesis test described later in this section.

The average rupture width of the 1700 Cascadia earthquake \( W = 74 \) km comes from the 3D dislocation model that best fits the tsunami heights measured in Japan (Satake et al., 2003). Using these same tsunami data, Satake et al. (1996) estimated the magnitude of the Cascadia earthquake to be 9.0 (8.8–9.2). Although the authors found that estimated tsunami heights were not that sensitive to using a wider width \( W = 100 \) km due to its trade-off with slip amplitude, they did find that the smaller width was needed to match the locations where subsidence occurred along the Pacific Northwest coast. The only other direct evidence that can be used to constrain the down-dip limit of rupture in the Cascadia earthquake is the amplitude of coastal subsidence. However, Wang et al. (2003) also found that there was a trade-off between \( W \) and slip amplitude in their 3D dislocation model of the Cascadia earthquake. Furthermore, they suggested that the amount of “coseismic” subsidence could be contaminated by post-seismic deformation and, therefore, overestimated.

Hyndman (2013) and Wang and Tréhu (2016) provide extensive summaries of all the available data that can be used to constrain the down-dip extent of coseismic rupture during great Cascadia earthquakes. Hyndman (2013) concluded that three of these data sources (i.e. geodetic, coastal paleoseismic, and thermal) agree with one another and with the reference 3D dislocation model of Flück et al. (1997). Flück et al. (1997) found that the locked zone of the subduction interface where strain is accumulating and full slip is expected to occur in a future great earthquake has an average width of 60 km with an additional 60 km average width of the transition zone in which slip is expected to decrease from full slip to zero. Wang et al. (2013) provide strong evidence that a conservative estimate of the point where slip goes to zero is about half way into the transition zone, giving a total average rupture width of \( W < 90 \) km. Rupture of the locked zone by itself provides a minimum estimate of rupture width. The estimate of Satake et al. (2003) falls in the middle of this range. Similar values of average rupture width for mega-earthquakes on the CSZ of 66 km (Priest et al., 2014) and 83 km (Witter et al., 2013) have been used for tsunami modeling. Wirth and Frankel (2019), using 3D ground motion simulations, found that coseismic rupture that extended to an assumed 1 cm/yr locking depth near the middle of the transition zone fit the coastal subsidence data better than down-dip rupture widths extending to the midpoint of the fully locked and 1 cm/yr locking depth or to the top of the episodic tremor and slip (ETS) events. Therefore, I believe that \( W = 74 \) (60–90) km is a realistic estimate of average rupture width for the 1700 Cascadia earthquake.

According to Table 3, the mean aspect ratios for the mega-earthquakes range from 2 to 7 (upper range of 3–9) excluding the Cascadia earthquake and up to 15 including it. The Cascadia earthquake is unique among the events because of its very long rupture length of 1100 km and relatively narrow mean rupture width of 74 km, although this is shown to be scientifically defensible in a later section of this report. The mega-earthquake magnitudes in Table 3 exceed the corresponding mean breakpoint magnitudes by an average value of 0.6 (0.2–1.3), consistent with their large aspect ratios. There is also significant overlap of
the ranges of $W$ and $W_{\text{MAX}}$ for all of the events. The earthquakes have an average rupture width of $W = 156 \pm 51$ km compared to an average mean seismogenic interface width of $W_{\text{MAX}} = 151 \pm 55$ km. A two-tailed $t$-test assuming unequal variances indicates that the hypothesis that the two means are different can be rejected at a 1% significance level. This is true even if the larger Nias-Simeulue rupture width is used and whether the relatively narrow Cascadia rupture width is excluded by itself or in conjunction with the larger Nias-Simeulue width.

The seven earthquakes with the largest rupture widths have widths of $181 \pm 36$ km, consistent with the $\approx 200$ km upper-bound rupture width suggested by Sea16, AH17, and Tajima et al. (2013). All of the comparisons presented in this section support the hypothesis proposed in this study that mega-earthquake rupture widths saturate consistent with a break in self-similar source scaling due to the physical limit of seismogenic interface widths, and by analogy that MSR reduces accordingly.

The results in Table 3 are also consistent with the proclivity of these subduction zones to produce both $M \geq 8.5$ mega-earthquakes, or so-called runaway (large aspect ratio) earthquakes (Woo and Mignan, 2018), according to the studies of Schellart and Rawlinson (2013) and Bletery et al. (2016). Furthermore, the mean breakpoint magnitudes are consistent with a $M \approx 8.5$ break in self-similar scaling of mega-earthquakes that has been proposed by many investigators cited in this article based on their observations and modeling of fault-width saturation of inferred earthquake ruptures.

**Comparison with 1700 Cascadia earthquake**

The 1700 Cascadia earthquake provides another consistency check of the hypothesis proposed in this study. The CSZ has a relatively small preferred seismogenic width of 68 km (Berryman et al., 2015), corresponding to a relatively small breakpoint magnitude of $M_B = 7.7 \pm 0.25$. However, there was a $M \approx 9$ (8.8–9.2) mega-earthquake on this subduction zone in 1700 that likely ruptured its entire length (Satake et al., 1996, 2003). The scaling relations referenced in this article can be used to reconcile the large difference between these magnitudes to see if such a small breakpoint magnitude is reasonable. If I assume a mean aspect ratio of 1.5 for an event corresponding to the breakpoint magnitude on the CSZ, I get a rupture length of about 100 km for a mean interface width of 68 km from the scaling relations listed in Table 1. Given a total rupture length of 1100 km (Satake et al., 2003), this corresponds to a factor of 11 increase in $A$ and $L$ because of the 1D rupture propagation. If I also assume that average slip scales according to the $L$-model (Scholz, 1982a), for which $D \approx L$, then slip also increases by a factor of 11 for a combined 121-fold increase in seismic moment according to Equation 1. Based on Equation 3, this 121-fold increase in seismic moment corresponds to a 1.4 increase in moment magnitude, which when added to $M_B = 7.7$ gives 9.1, within the range of the estimated magnitude of the 1700 earthquake.

The estimated value of $D$ for the 1700 earthquake is 14 m (Satake et al., 2003). Assuming $L$-model scaling, this would correspond to an average slip of about 1.3 m for a $M$ 7.7 event, which is within the range of median estimates of $D$ from all the scaling relations cited in this article, confirming the reasonableness of the estimated $M_B = 7.7$ for the CSZ using the methodology proposed in this study. A possible reason for such a large discrepancy between the breakpoint magnitude and the magnitude of the 1700 earthquake is presented in a later section of this article.
**Comparison with empirical GMMs**

The best check of whether the breakpoint magnitudes based on fault-width saturation are consistent with a break in MSR is to compare them with the empirical subduction GMM of Abrahamson et al. (2018), which is the only current one that uses an up-to-date global subduction earthquake database. Using the NGA-Subduction database described by Kashida et al. (2018), Abrahamson et al. (2018) found that it was necessary to reduce the MSR of short-period ($T < 0.75$ s) ground motions of megathrust events for $M > 8.2$ (decreasing to 7.8 at $T \geq 5$ s) to accommodate both the Maule and the Tohoku-Oki ground motion data and MSRs predicted by ground motion simulations of megathrust earthquakes (e.g. Atkinson and Macias, 2009; Gregor et al., 2002, 2006). These breakpoint magnitudes are 0.2 higher than those of Abrahamson et al. (2016), which it supersedes. Unfortunately, the Maule and Tohoku-Oki earthquakes are the only events in the NGA-Subduction database with magnitudes above the estimated breakpoint magnitude of the host subduction zone, which makes both the breakpoint magnitude and MSR uncertain. Nevertheless, the breakpoint magnitudes determined in this study averaged overall all subduction zones ($7.9 \pm 0.26$ from Table 2) are generally consistent with the break in the MSR in the GMM of Abrahamson et al. (2018).

I surmise that the smaller MSR breakpoint magnitudes at longer periods found empirically might be due to differences in dynamic propagation characteristics between high- and low-frequency ground motions. Low-frequency ground motions scale more similarly to static displacement $D$ (e.g. Zhu and Rivera, 2002), which for the $L$-model is proportional to $L$. They also are less impacted by heterogeneities on the rupture surface and along the propagation path. However, high-frequency ground motions are more impacted by small-scale heterogeneities (asperities, barriers, roughness) on the subduction interface and along the propagation path, leading to greater intrinsic damping, scattering, and frictional heating and a correspondingly lower MSR than that of low-frequency motions. The larger MSR at low frequencies might lead to a slightly lower apparent breakpoint magnitude intersection point because of its steeper slope. Further studies will be required to gain a better understanding of this phenomenon. A similar frequency-dependent MSR for large megathrust earthquakes has been found by Atkinson and Boore (2003, 2009), Gregor et al. (2002, 2006), Ghofrani and Atkinson (2014), Abrahamson et al. (2016, 2018), and Zhao et al. (2016), making it a pervasive attribute of ground motion scaling.

**Comparison with intraslab breakpoint magnitudes**

Another consistency check of the interface breakpoint magnitudes proposed in this study is to compare them with MSR breakpoint magnitudes found by Ji and Archuleta (2018) for subduction intraslab earthquakes. Slab thicknesses are significantly less than seismogenic interface widths for an individual subduction zone because of the shallow dips of the slabs. Even accounting for the sub-horizontal and sub-vertical dip angles of the simulated intraslab events, the maximum allowable rupture widths are less than the interface widths. Therefore, I would expect megathrust breakpoint magnitudes estimated in this study to be greater than MSR magnitude breakpoints estimated for intraslab earthquakes. For this comparison, I used the range of “breakoff” and “plate-unbending” saturation magnitudes given in Ji and Archuleta (2018) as recommended by Chen Ji (personal communication, 2018). For example, the CSZ has an interface breakpoint magnitude of $7.7 \pm 0.25$ and an intraslab breakpoint magnitude ranging from 6.3 to 7.1; the Japan Subduction Zone (Tohoku region) has an interface breakpoint magnitude of $8.5 \pm 0.28$ and an intraslab
breakpoint magnitude ranging from 7.6 to 8.2; and the South America Subduction Zone (Central Chile region) has an interface breakpoint magnitude of 8.7±0.30 and an intraslab breakpoint magnitude ranging from 7.1 to 8.0. Similar differences of ≈ 0.5 are found for the other subduction zones listed in Ji and Archuleta (2018), consistent with my expectation that the breakpoint magnitudes based on seismogenic interface width (a proxy for a break in MSR) should be larger than those for intraslab earthquakes.

Ji and Archuleta (2018) did not provide empirical evidence to support their proposed breakpoint magnitudes for intraslab earthquakes and doing so is beyond the scope of this study. However, I note that Abrahamson et al. (2016) and Zhao et al. (2016) used breakpoint magnitudes of 7.5 and 7.1, respectively, for intraslab events in their GMMs, although they too had little data to support them. These breakpoint magnitudes are generally consistent with the slab unbending magnitudes of the majority of the subduction zones with relatively large slab thicknesses studied by Ji and Archuleta (2018). Both Sea10 and AH17, the latter using source dimensions from the consistently derived finite-source inversions described in Hayes et al. (2015) and Hayes (2017), proposed linear scaling relationships between log(W) and M for intraslab events, although AH17 suggested that single-rupture intraslab earthquakes are generally not expected to grow beyond M 8.0 because their rupture zones are limited by the thickness of the subducting slab, similar to the hypothesis of Ji and Archuleta (2018). This breakpoint magnitude is generally consistent with the slab break-off magnitudes of the majority of the subduction zones with relatively large slab thicknesses studied by Ji and Archuleta (2018). The large range of rupture widths reported by Hayes (2017) and used by AH17 (25–120 km) between M 7.5 and 8.0 led these latter authors to adopt the slope derived from interface events for their intraslab log(W) versus M relationship.

One recent intraslab event that many seismologists believe shows rupture width saturation is the M 8.1–8.2 Chiapas (Tehuantepec) earthquake of 8 September 2017 that occurred along the Middle America Subduction Zone off the coast of southern Mexico. Source inversions and aftershocks indicate that the rupture width of this earthquake was largely confined to the 30-km-thick Cocos slab and had an aspect ratio of ≈ 2–4 (e.g. Chen et al., 2018; Jiménez, 2018; Okuwaki and Yagi, 2017; Suárez et al., 2019; Ye et al., 2017). Ji and Archuleta (2018) estimated M_b = 7.5 – 7.7 for the slab break-off (slab detachment) scenario for this subduction zone similar to the scenario that has been proposed for the Chiapas earthquake by seismologists. In this case, the earthquake’s large magnitude comes from its relatively long rupture length and large slip. However, some seismologists (e.g. Melgar et al., 2018; Zhang and Brudzinski, 2019) proposed an alternative scenario in which the Chiapas earthquake ruptured the entire oceanic lithosphere (brittle slab plus mantle). Nevertheless, most studies agree that the largest slip and majority of seismic moment was released within the oceanic slab, which dominated both the seismic moment and the strong ground motion for the earthquake. Furthermore, none of the slip models have yet to be trimmed to eliminate spurious effects due to the relatively noisy and non-unique inversion process as recommended by Hayes (2017).

Sahakian et al. (2018) evaluated the recorded values of PGA from the Chiapas earthquake and found that they were consistent out to a distance of 400 km with the magnitude-extrapolated predictions for intraslab earthquakes from the GMMs of Garcia et al. (2005), Zhao et al. (2006), and Abrahamson et al. (2016). In fact, these investigators found that the recorded PGA values within about 200 km were actually more consistent with the predictions from the subduction interface GMM of Zhao et al. (2006) than with those from the intraslab GMMs. This latter result together with the fact that all of the
GMMs had terms to account for near-source distance saturation and magnitude saturation supports the hypothesis of Ji and Archuleta (2018), applied herein to megathrust earthquakes, that the MSR of ground motion reduces beyond some breakpoint magnitude that is related to the maximum seismogenic width of the fault.

Application to the CSZ

Berryman et al. (2015) estimated a relatively narrow preferred average seismogenic interface width of $W_{MAX}$ = 68 km for the CSZ based primarily on the most likely down-dip limit of coseismic rupture defined by the midpoint of the transition zone estimated from geodetic and thermal data (Frankel and Petersen, 2008). This resulted in a relatively small estimate of breakpoint magnitude ($M_B = 7.7^{+0.5}_{-0.4}$). A similar interface width ($W_{MAX}$ = 64 km) and breakpoint magnitude ($M_B = 7.7^{+0.4}_{-0.4}$) was derived for the Nankai segment of the Nankai-Ryukyu Subduction Zone by Berryman et al. (2015), which is consistent with it serving as an analogy to the CSZ due to its similar age of subducting oceanic crust (~10 million years old) and rate of subduction (4–8 cm/yr) (Satake and Atwater, 2007). Furthermore, Frankel et al. (2018) found that their broadband ground motion simulations of a $M$ = 9 mega-earthquake on the CSZ were consistent with predictions from the empirical GMM of Abrahamson et al. (2016), which has a central breakpoint magnitude of $8.0^{+0.2}_{-0.2}$ at short periods decreasing to $7.6^{+0.2}_{-0.2}$ at long periods, similar to the mean estimate proposed in this study and within its epistemic uncertainty bounds. I also showed in a previous section of this report that the preferred seismogenic width and resulting breakpoint magnitude for the CSZ are consistent with the average magnitude and slip estimated by Satake et al. (2003) for the 1700 Cascadia earthquake assuming $L$-model scaling.

The Cascadia subduction interface shallows in dip and widens considerably in the Puget Sound region where the North American plate bends to a more northwesterly direction. The most likely seismogenic interface width in this region is ~140 km according to Frankel and Petersen (2008), although Frankel et al. (2014) revised this to ~170 km based on the inferred 1 cm/yr locking depth recommended at a regional workshop of experts convened as part of the National Seismic Hazard Model Program. The $W_{MAX}$ ~ 200 km upper range of seismogenic width used by Frankel and Petersen (2008) might be considered an upper bound for earthquakes that initiate in the Puget Sound region given that it represents a rupture width extending to the top of ETS events, whereas both Hyndman (2013) and Wirth and Frankel (2019) showed that coseismic rupture to this depth is not likely given thermal and other constraints.

The question is whether an earthquake that originates on the wider section of the CSZ interface in the Puget Sound region will exhibit a larger breakpoint magnitude. This might be the case, for example, if rupture were to extend the full width of the fault before extending along strike rather than rupturing along strike before “filling in” the full width of the seismogenic interface in this region. If the first hypothesis is the case, one might expect a similar breakpoint magnitude to the Andaman segment of the Andaman-Sunda Subduction Zone, for which $W_{MAX}$ = 143 km and $M_B = 8.4^{+0.6}_{-0.4}$, or to the Japan segment of the Japan-Kuril subduction zone, for which $W_{MAX}$ = 166 km and $M_B = 8.5^{+0.7}_{-0.4}$. To account for this uncertainty, a larger breakpoint magnitude could be used for hypocenters located in the Puget Sound region and a smaller one (e.g. Table 1) for hypocenters originating along the narrower sections of the CSZ. Alternatively, a single set of breakpoint magnitudes could be used regardless of hypocenter location that spans the above values. For example, one possible set of values would be $M_B = 7.7$ (5% CL), 8.0 (mean), and 8.5 (95%
Conclusion

Based on the studies reviewed in this article, I propose a methodology for estimating the breakpoint magnitude at which subduction megathrust earthquakes exhibit a breakdown in self-similar scaling of both source dimensions and ground motion once their rupture width saturates the seismogenic width of the subduction interface and rupture transitions from 2D to 1D scaling. Although empirical evidence is sparse, I present several seismological studies that support such a breakdown for megathrust earthquakes. The concept of a breakpoint magnitude is also supported by ground motion simulations of subduction intraslab events and the large body of empirical and seismological evidence of a similar breakdown in source-dimension scaling and MSR for shallow continental earthquakes. Using epistemic uncertainty in seismogenic interface widths, source scaling relations, and rupture aspect ratios, I use the proposed methodology to estimate the weighted epistemic mean, median, standard deviation, 5% CL, and 95% CL of the breakpoint magnitudes for 79 global subduction zones, including Cascadia, defined in Berryman et al. (2015). Cascadia is a special case because of its varying interface widths that indicates that the breakpoint magnitude might be different depending on the location of the hypocenter of a future Cascadia event. Other subduction zones might have similarly varying interface widths but identifying them was beyond the scope of this study.

The breakpoint magnitudes derived in this study can be used to constrain the magnitude at which MSR is expected to decrease (i.e. the slope of MSR becomes shallower) for a specific subduction zone when predicting ground motions from existing and future subduction GMMs. This can have large consequences on the prediction of ground motion amplitudes for magnitudes larger than the breakpoint magnitude. However, there are too few mega-earthquakes with sufficient recordings to empirically determine what this MSR should be, so one must look to the limited number of ground motion simulations to help constrain it. More ground motion simulations of megathrust earthquakes are also needed to further confirm the relationship between interface width and the break in MSR and the inferred break in source scaling at $M \approx 8.5$ that has been suggested for megathrust earthquakes by many seismologists. Therefore, a potential user of the breakpoint magnitudes derived in this study should use judgment in selecting appropriate values to use depending on the recorded magnitudes and material properties that are available for the specific subduction zone of interest.

Acknowledgments

The motivation for this study came from a similar concept proposed for subduction intraslab earthquakes by Chen Ji and Ralph Archuleta. I want to thank Yousef Bozorgnia, Nico Kuehn, and Nick Gregor as well as other participants of the NGA-Subduction project for their support and comments throughout the study. I also would like to recognize three anonymous reviewers for their thorough and insightful comments on how to improve the manuscript.

Declaration of conflicting interests

The author(s) declared no potential conflicts of interest with respect to the research, authorship, and/or publication of this article.
Funding
The author(s) disclosed receipt of the following financial support for the research, authorship, and/or publication of this article: CoreLogic, Inc. will pay for the publication expenses of this article.

References


