Superquakes and Supercycles

Chris Goldfinger¹, Yasutaka Ikeda², Robert S. Yeats³, Junjie Ren⁴

¹Oregon State University, College of Oceanic and Atmospheric Sciences, 104 Ocean Admin. Bldg., Corvallis OR 97333-5506, USA. gold@coas.oregonstate.edu

²Department of Earth and Planetary Science, University of Tokyo (Hongo 7-3-1, Bunkyo-ku, Tokyo 113-0033, Japan. ikeda@eps.s.u-tokyo.ac.jp

³Oregon State University, Department of Geosciences, 104 Wilkinson Hall, Oregon State University, Corvallis, OR 97331-5506, USA. yeatsr@geo.oregonstate.edu

⁴Key Laboratory of Crustal Dynamics, Institute of Crustal Dynamics, China Earthquake Administration, Beijing 100085, China. renjunjie@gmail.com

Online material: Location maps, core data, radiocarbon data, additional figures with individual plots for megathrust energy state by site, energy state for NE Japan, event parameters and correlation matrix.
http://activetectonics.coas.oregonstate.edu/SRL/esupp-Goldfinger et al.html

INTRODUCTION

The recent 2011 Mw=9.0 Tohoku Japan, and the 2004 Mw=9.2 Sumatra-Andaman superquakes have humbled many in earthquake research. Neither region was thought capable of earthquakes exceeding Mw~8.4. Appealing proposed relationships to predict the size of earthquakes in subduction zones such as that between earthquake magnitude and parameters such as lower plate age and convergence rate (Ruff and Kanamori, 1980) and plate coupling based on anchored slabs (Scholz and Campos, 1995), at least have many exceptions, and may not be valid. Both earthquakes occurred where the plate age was quite old, ~ 50-130 my. The role of thick sediments smoothing the plate interface and maximizing rupture area has been considered a contributing factor, and seems influence many recent great earthquakes (Ruff, 1989). The Tohoku event is also contrary to this hypothesis. Clearly much remains to be learned about these great events, so much so that most previous estimates of maximum earthquake size in subduction
plate boundaries should be considered suspect, and perhaps other fault systems as well (McCaffrey, 2007, 2008).

Our perspective on this issue is clearly hampered by short historical and even shorter instrumental records. The examples noted above indicate that basing estimations of maximum earthquake size or models of earthquake recurrence on such short-term records alone clearly cannot encompass the range of fault behavior, even when historical records may be > 1000 years long as in Japan. Here we present several examples of areas where long geologic and paleoseismic records can illuminate a much wider range of seismic behaviors than those deduced from historical and instrumental data, and speculate on models of long-term fault behavior based on very long records.

**NORTHEAST JAPAN TRENCH**

Since Pliocene time, the Japan arc has been subjected to east-west compression due principally to the westward convergence of the Pacific plate at the Japan Trench at a rate of 70 mm/yr. (DeMets et al., 2010). Normal faults, likely resulting from Miocene back-arc spreading, have been reactivated in Plio-Quaternary time as thrust faults due to the change in regional stress fields from tension to E-W compression.

Geologic constraints indicate this upper-plate compression is only a small fraction of the overall total convergence between Japan and the Pacific plate. Pliocene-Quaternary deformation is concentrated particularly within the Uetsu fold-and-thrust belt and its southern extension, the Northern Fossa Magna, which extend along the Japan Sea coast (Sato, 1989). This fold-and-
thrust belt develops within a rifted basin formed in middle Miocene time in conjunction with the opening of the Japan Sea (Sato, 1994). The rate of horizontal shortening over the Uetsu fold-and-thrust belt in Pliocene to Quaternary time has been determined to be 3-5 mm/yr. (Okada and Ikeda, 2012). Including other active faults and folds, the total rate of horizontal shortening over the Northeast Japan arc is estimated to be 5-7 mm/yr, in good agreement with previous estimates (Wesnousky et al., 1982). Deformation to the west in the Sea of Japan is thought to be minor based on marine seismic reflection data and seismicity (Okada and Ikeda, 2012) and GPS data that suggests little relative motion between the west coast of NE Japan and stable Eurasia (Sagiya et al. 2000). Inelastic deformation of the submarine forearc is weakly extensional for the most part (i.e von Huene and Lallemand, 1990). This indicates that only a fraction (< 10%) of plate convergence is accommodated in the Japan arc as inelastic, permanent deformation. The geologically inferred uplift rate averaged over the Northeast Japan arc in late Quaternary time is 0.22-0.36 mm/yr. (Tajikara, 2004). If the Northeast Japan arc is in isostatic balance, then regional uplift/subsidence occurs due to (1) crustal thickening, which in turn is caused by crustal shortening and magmatic underplating/intrusion, (2) surface unloading by denudation, and (3) surface loading resulted from deposition of volcanic and eolian materials. The total rate of crustal shortening was determined from the regional uplift data by eliminating the other disturbing factors, and was found to be < 6-8 mm/yr. (Hashimoto, 1990), in good agreement with the shortening rate deduced from active fault and fold data.

The geodetically observed, short-term deformation rate within the Northeast Japan arc is, however, significantly larger than the long-term deformation. Triangulation, trilateration and GPS observations during the last ~100 years revealed that the Japan arc has contracted in an east-
west direction at a rate as high as several tens of mm/yr. (Hashimoto, 1990; Suwa et al., 2006; Sagiya et al., 2000). This rate is nearly one order of magnitude greater than geologically observed shortening rates, and is comparable to the ~ 83 mm/yr rate of plate convergence at the Japan Trench (Fig. 1).

Similarly, there is a contrast between recent rapid coastal subsidence and the long-term evidence of terrace uplift along the Pacific coast of NE Japan. Tide gauge data indicate abnormally high rates (several to 10 mm/yr) of subsidence during the last ~80 years (Kato, 1983). This subsidence is likely due to strong coupling dragging down the upper plate by the subducting Pacific plate beneath the Japan arc. However, late Quaternary marine terraces developing along the Pacific coast indicate uplift at 0.1-0.4 mm/yr. (Koike and Machida, 2001). The discrepancy between short-term (geodetic) and long-term (geologic) observations indicates that most of the strain accumulating in the last ~100 years has been elastic, to be released by slip in large earthquakes on the subducting plate boundary. Only a fraction of the strain would remain in the arc as inelastic deformation. Although large thrust-type earthquakes with magnitude 7-8.4 have occurred at the Japan Trench during the last 100 years, they did not result in significant strain release on land. Thus, larger slip events are required to occur at intervals much longer than the period (~100 years) of instrumental observations (Ikeda, 2003, 2005). The recent Tohoku-oki Mw =9.0 earthquake was such a slip event with a slip based on submarine GPS of~ 24m (Sato, 2011), and slips greater than 50 m near the toe of the accretionary prism (e.g. Fujiwara et al., 2012); its rupture area encompassed those of numerous previous earthquakes of magnitude 7-8.4. A deficit of slip was also inferred in this region from sequences of small repeating earthquakes (Igarashi, 2010). The fact that both the smaller earthquakes and the 2011 earthquake all caused
coastal subsidence as well, is likely related to deeper coupling that was not released
coseismically (Ikeda, 2012).

The most recent earthquakes in the Sendai area: 1933 (M 8.1), 1936 (M 7.5), and 1978 (Mw 7.6)
did not leave a tsunami record in nearby Suijin-numa, a coastal lake (Sawai et al., 2008), nor did
they leave extensive sand sheets on the Sendai Plain (Minoura et al., 2001). However, the older
historical record in the Tohoku region in Japan includes a number of large earthquakes and
associated tsunami, including large events in 869, 1611 and 1896. The largest of these events
was likely the 869 Jogan tsunami based on the presence of tsunami deposits in the coastal lake
(Sawai et al., 2008), the 3-4 km landward extent of inundation relative to the paleo-shoreline
(Sawai et al., 2007; Shishikura et al., 2008; Sugawara et al., 2012) and tsunami modeling
(Namegaya et al., 2010). The landward inundation of the March 2011 tsunami similarly
penetrated 3-4 km inland on the Sendai plain (http://earthobservatory.nasa.gov). The
paleotsunami evidence also includes two predecessors to the Jogan event that also penetrated ~ 4
km inland (though in Jogan time the shoreline was ~ 1 km west of present). These large tsunami
support the existence of periodic outsized earthquakes (Mw~ 9) along the Tohoku coast. The
recurrence times are between 800 and1200 years, with numerous smaller events between that
make up the majority of the historical record (Sawai et al., 2008; Shishikura et al., 2007). Our
field transect across the Ishinomaki Plain verifies the result of Shishikura, and suggests that the
Jogan tsunami deposits are locally more robust even that the March 2011 tsunami deposit, even
accounting for coastline change, supporting the inference that Jogan may have been and M~ 9
earthquake (e-supp. Fig. S1).
In Hokkaido, a similar relationship has been observed between the shorter historical and instrumental records, and the paleoseismic record along the northern Japan Trench. Prehistoric tsunami that most likely were generated from long ruptures along the northern Japan Trench were significantly larger than those generated by earthquakes of Mw=7-8.3 in the historical and instrumental records (Nanayama et al., 2003). Over the past 2000-7000 years, such outsized events (M~ 9) occurred on average about every 500 years in the Hokkaido region, with the most recent event ~ 350 years ago (Nanayama et al., 2003).

HIMALAYAN FRONT AND HAIYUAN FAULT

The Himalayan front region experienced four earthquakes in the period 1897-1950 with magnitudes of 7.8 to 8.4 that have been assumed to be the MCE (Maximum Considered Earthquake), or design earthquake, for power plants and large hydroelectric dams. However, surface rupture accompanying these four historic earthquakes was either absent or small (summarized by Yeats, 2012). On the other hand, paleoseismological evidence exists for surface displacements as large as 26 m from an earthquake in the late 15th-early 16th century along the Himalayan front of northwest India (Kumar et al., 2006) that may correspond to an earthquake in 1505 A.D. (Iyengar et al., 1999). An earlier earthquake around 1100-1200 A.D. in Nepal and adjacent Arunachal Pradesh state in northeast India (Fig. 2) was accompanied by a single-event surface rupture of 17m or larger. Rupture lengths accompanying these earthquakes may be possibly the largest known worldwide on a continental reverse-fault earthquake (Lavé et al., 2005; Kumar et al., 2010; Fig. 2). Trenching of this fault supports a large single rupture, as opposed to several smaller events closely spaced in time. It has been suggested that the rupture length for both the 1505 A.D. and the 1100-1200 A.D. earthquakes might be close to 900 km (Kumar et al., 2010; Yeats, 2012). A 900 km rupture length implies magnitudes greater than 8.5 (Wells and
Coppersmith, 1994); however, uncertainty exists in the exact timing of these earthquakes, due to uncertainty in correlating dates in trenches with historical events in both India and Nepal.

Another example of the contrast in inferred MCE based on recent historic information with longer-term geologic data is found along the Haiyuan fault in Gansu Province, China. This fault was the source of an earthquake in 1920 with a rupture length of 237 km that killed more than 220,000 people (e-supp. Fig. S2). An extensive paleoseismic trenching program along the entire length of the fault showed that the 1920 earthquake was not typical of this fault. Yonkang et al. (1997) divided the 1920 rupture into three segments and dated surface-rupturing earthquakes in each segment over the past 6000 yrs. They found that some earthquakes ruptured one segment, and some ruptured two, but only one pre-historic earthquake (6100-6200 yrs. BP) ruptured all three segments and may have been a duplicate of the 1920 event (Yongkang et al., 1997). In this example, the most recent earthquake was the largest, and prior to trenching, there was a tendency to regard the 1920 earthquake as the typical or characteristic earthquake. The paleoseismic evidence showed that the majority of the earthquakes were much smaller. Supplementary table 2 also shows that the two largest events had much greater net slip (5.6 and 7.0 m respectively for the ~6150 BP and AD 1920 events) than the 1.5-2 meter average for seven intervening single segment ruptures (6 additional single and one three segment rupture could not be determined) (State Seismological Bureau, 1990).

CASCADIA

In Cascadia, several decades of paleoseismic work have yielded an unprecedented record of great earthquakes. Pioneering work (Atwater, 1987) established the repeated occurrence of great
earthquakes and tsunami along the Washington coast, followed by widespread evidence of subsidence and tsunami from coastal sediments along the entire margin from Canada to northern California (Atwater et al., 1997; 2004, Kelsey et al., 2005; Clague and Brobowski, 1994). The longest records available, those from deep-sea turbidites, reveal the complex behavior of this subduction zone over the past 10,000 years (Goldfinger et al., 2012; e-supp. Fig. S3). The offshore records are in good agreement with onshore paleoseismology where temporal overlap exists (variable from 3500-4600 years BP), and both offer consistent information about the relative size of paleoearthquakes, giving confidence that both are recording the same phenomenon (Goldfinger et al., 2012). The longer records reveal several important features not discernible with shorter records. One is that there is apparent clustering of the larger (rupture lengths greater than 600 km, Mw~ 8.7-9.0) events into groups of 4-5 events, with 700-1200 year gaps between the clusters (Goldfinger et al., 2012). Another is significant segmentation of the margin, with a group of shorter ruptures limited to southern Cascadia that are interspersed between the long ruptures as determined by intersite stratigraphic correlation (Goldfinger et al., 2008; 2012 their Fig. 55).

Evidence of the relative size of these great events comes from comparisons of turbidite thickness and rupture length using the extensive spatial sampling of the cores (e-supp. Fig. S3), and suggests that there may be earthquakes larger than the well known A.D. 1700 event. Goldfinger et al. (2012) show that mass per event down core among four key Cascadia core sites is reasonably consistent among sites. They infer that the best explanation is that turbidite thickness and mass are linked to the relative levels of ground shaking in the source earthquakes. The magnitude of the AD 1700 earthquake is estimated to be ~ 9.0 based on tsunami inversion of the
tsunami heights along the Japanese coast, and the attribution of this “orphan tsunami” to Cascadia (Satake et al., 2003). The magnitude could change in the future with more sophisticated modeling, but provides a benchmark for other paleoseismic events. The connection between earthquake size and turbidite size is tenuous; nevertheless, the correlation between size characteristics per event among numerous cores along strike strongly suggests a regional connection that can best be attributed to the magnitude or shaking intensity of the source earthquake (Fig. 3, Goldfinger et al., 2008, 2012). In the offshore turbidite record, the turbidite associated with the 1700 AD event is roughly “average” in mass and thickness relative to 19 inferred similar ruptures as compared between core sites along the 1000 km Cascadia margin (Goldfinger et al., 2012). Significantly, there are several turbidites that are considerably larger in terms of thickness and mass in the 10,000 year offshore record. Notable are the 11th and 16th events back in time, known as T11 and T16, that took place 5960 +/- 140 and 8810 +/- 160 years ago (Fig. 3). These two turbidites are consistently larger at all core sites along the length of Cascadia (e-supp. Fig. S4 and Table S3), being an average of 2.9 times (range 2.8-3.1) the A.D. 1700 turbidite mass. At most sites, 4-7 other events in the 10ka turbidite record, (typically including turbidites T5, T6, T7, T8, T9, T13 and T18) are also larger than the 1700 AD turbidite, though by smaller margins of 1.3-1.7 times the T1 turbidite mass (range 0.2-1.7). Goldfinger et al. (2012) estimate Mw for all 19 Cascadia ruptures of 600 km and greater using estimated rupture length, width and slip parameters calibrated to the AD 1700 event, and setting that event equal to 9.0. They estimate Mw for T11 and T16 to be ~9.1. The average mass increase in these two turbidites of 2.8 times the AD. 1700 turbidite is roughly comparable to the energy increase of 1.4 times from Mw 9.0 to Mw 9.1. The event to event variability and consistency among sites are unlikely to be due to changes in sediment supply, oceanography, or other factors as they are
replicated at numerous sites, including one (Hydrate Ridge) with no modern terriginous sediment supply. The outsized turbidites are unlikely to be due to a long prior sediment accumulation interval, as only T11 has such a prior interval (~ 1100 years interrupted by one small event), and because other events following ~ 1000 year gaps were not outsized in thickness or mass (T6 for example). These data suggest that significantly outsized earthquakes may occur at a rate of 1-2/10,000 years in Cascadia.

DISCUSSION

Uniquely, the 10 ka Holocene Cascadia earthquake generated turbidite stratigraphy affords uncommon opportunities to examine recurrence models, clustering and detailed long-term (10 ka) strain history of a subduction zone. First, in Cascadia, the two outsized superquakes do not appear to occur in an otherwise random sequence. Cascadia earthquakes appear to cluster, with the larger events that include much of the strike length of the margin occurring within groupings of 4-5 events that comprise four Holocene clusters. There appears to be a weak tendency to terminate these clusters with an outsized event. The long time series also suggests that Cascadia is neither time nor slip predictable (Goldfinger et al., 2012).

Because there appears to be a connection between earthquake size and turbidite size among core sites and across a variety of depositional environments, an opportunity exists to investigate the earthquake pattern further. This inference comes from the observation that correlated turbidites along strike in Cascadia vary considerably in mass and thickness per event at each site in the Holocene series, but that they are consistent in mass and thickness for the same event at multiple sites and multiple depositional environments as previously described (Goldfinger et al., 2012; e-supp. Fig. S4, Tables S4, S5). Because of this consistency for
individual events along the margin, and despite the obvious simplifications involved, we infer that turbidite mass can be considered a crude proxy for seismic moment or intensity of ground shaking at offshore sites for at least the 19 larger ruptures of 600 km or greater described in Goldfinger et al. (2012).

If our assumption that energy release can be approximated by turbidite mass and thickness, we can then assemble and compare the Holocene series of earthquakes as a time series. First, we assume that while slip and moment of paleoearthquakes is unknown, that coseismic energy release may be modeled as proportional to the mass of turbidites triggered in seismic shaking. Second we assume that plate convergence between earthquakes increases elastic strain energy in proportion to interevent time (a coupling coefficient of 1.0 is assumed).

To examine the energy balance between subduction earthquakes and accumulation of elastic strain, we scale turbidite mass (energy release) to balance plate convergence (energy gain) to generate a 10ka energy time series for Cascadia (Fig. 4). We do not know the starting or ending values of course, thus we simply scale the plot such that the overall trend of the series has no net gain or loss of potential energy. The interval between the last earthquake and the next one is also unknown, and we set this equal to the average recurrence time. Sources of error also include the uncertainties in the radiocarbon ages for each event, which are taken from Goldfinger et al. (2012) and are shown on the plot. Both approximations undoubtedly comprise additional sources of error. The resulting sawtooth pattern reveals what we interpret as a complex pattern of long-term energy cycling on the Cascadia megathrust, with the vertical scale representing potential energy. If correct, we can then make some observations about the long-term behavior of Cascadia. Earthquake clusters including small to large events appear to have significant variations in energy balance within and between the clusters. Cluster four (~10000-8800 BP)
appears to maintain a relatively even energy state comprising several seismic cycles before falling to a low after large event T16. Cluster three (~8200-5800 BP) climbs steadily in energy state through multiple seismic cycles until falling sharply to a similar low following large event T11. Cluster two (~4800-2500 BP) climbs then falls to a low energy value after T6, which also precedes a long gap of ~1000 years, which then raises the energy state. Cluster one (~1600-300 BP) slowly declines from T5 to T1, the AD 1700 Mw~ 9.0 earthquake.

Overall, what is suggested by this pattern is that some events release less energy while others release more energy than available from plate convergence (slip deficit) and may have borrowed stored energy from previous cycles. This suggests that energy release in the earthquakes is not closely tied to recurrence intervals, that is, they are not obviously slip or time predictable, but the pattern of values suggests that it is not likely to be a Poisson process either. The highest energy states may result in either a very large earthquake, or a series of smaller earthquakes to relieve stress. A very low energy state may result in a long gap or in a series of smaller earthquakes with a net energy gain over time, something that also appears to describe NE Japan prior to the 2011 Tohoku earthquake. Although the starting and ending points in Fig. 4 are unknown, and the scale factor is based only on the condition of no net energy change over 10,000 years, we note that changes in these three parameters cannot alter the pattern observed, nor make the series either time or slip predictable (e-supp. Figs. S5, S6). Neither would a different value for the coupling coefficient alter the pattern. If we sum the seismic moment for Cascadia using the geographic dimensions and rates in Goldfinger et al. (2012), the total seismic moment available is $6.7 \times 10^{30} \text{dyne/cm}$. If we then sum the inferred seismic moment from the individual events and their dimensions from the same source, the total is $6.1 \times 10^{30} \text{dyne/cm}$, implying a seismic coupling coefficient of ~ 0.9, though this is very approximate at best. The
seismic coupling coefficient, however, does not influence the result shown in Fig. 3 unless this value varies from event to event, a possibility that cannot be ruled out. It is also possible that the 10,000 year time series is not long enough to estimate this parameter. We include a similar figure showing the energy state in NE Japan, based on the paleoseismology available and historical earthquakes in the rupture zone of the 2011 earthquake (e-supp. Fig. S7).

Long-term “Supercycles” (Sieh et al., 2008) may help explain observed mismatches between deformation models based paleoseismology and those based on geodetic and other datasets. For example Bradley Lake, Oregon is a tsunami deposit site in a coastal lake that provides an excellent sensitivity test of the energy generated by past earthquakes. While the tsunami deposits in Bradley Lake are good temporal correlatives for offshore turbidites (Goldfinger et al., 2012), the fault slip required to generate tsunami that reach the lake appears inconsistent with the time intervals between events. Tsunami models generated using slip expected from plate motion for late Holocene earthquake ruptures in the most recent cluster are 170-340 years. Yet tsunami generated from full coupling of the North American and Juan de Fuca plates does not produce a large enough tsunami to reach the lake with these values (Witter et al., 2012). Moving all parameters in the direction of a larger tsunami (broader locked interface, full rupture in deep water at the trench), and including a splay fault (which Witter et al. 2012 did not believe was significant) still failed to generate sufficient inundation (Witter et al., 2012). This discrepancy may be explained by invoking a long-term cycle in which a series of earthquakes use long-term energy stored from previous seismic cycles (Goldfinger et al., 2010; Witter et al., 2012), possibly that some of the smaller events represent unbroken patches from previous heterogeneous ruptures (Witter et al., 2012).
CONCLUSIONS

Collectively, these data on the timing and size of geologically recorded earthquakes is not always consistent with the historic or instrumental record. It is becoming increasingly clear that our short instrumental and historical records are inadequate to characterize the complex and multi-scale seismic behavior of subduction zones and other major fault systems. The recurrence intervals for superquakes and supercycles may be very long, potentially biasing the record from the start. The data presented here suggests that the types and sizes of large earthquakes a given fault system is capable of producing may be unknown for most major fault systems. This too is a function of the short histories available. This bias toward what we have observed or recorded directly has shaped the development of conceptual models such as the characteristic earthquake model, the seismic gap theory, the relationship between plate age and convergence velocity, and time and slip-predictability.

Given recent failures of conceptual models of great earthquake recurrence, we must consider that there may not be reliable predictive factors available for such purposes at present, or in the immediate future, since many major fault systems lack both paleoseismology and GPS coverage. Evidence for deep coupling in NE Japan (Kato, 1983) and perhaps Cascadia (Chapman and Melbourne, 2009), suggest our knowledge of plate coupling is far from complete.

Paleoseismology offers a relatively simple method of determining the long-term behavior of a fault system, in the best of circumstances providing information on not just the time series of event occurrences and current strain accumulation, but segmentation, clustering, magnitudes, stress transfer. The evidence here for strain supercycles that transcend individual seismic cycling is troubling even for short-term paleoseismic records. The paleoseismic data however
can be helpful in eliminating seismic cycle scenarios that have never occurred, while emphasizing those that did.

In terms of seismic hazards, particularly for critical facilities, we suggest that characterization of the maximum considered earthquake, may require earthquake records over 10’s of past events, perhaps as long as 10,000 years. Credible earthquake scenarios that rely on instrumental or even relatively long historical records as in NE Japan, may or may not capture the range of earthquakes possible for a given fault system. Bias can occur in either direction, as the case of the 1920 M~7.8 Haiyuan China earthquake, which was larger than most of the earthquakes on that fault, while the 2011 Tohoku event seemed outsized relative to recent history. Long records put both types of bias in perspective.

“Superquakes” appear to result from linkups of fault segments that more commonly have separate histories. These segments may connect in long ruptures in ways and at times that cannot presently be predicted. They may also result in larger slips and possibly broader ruptures, entraining regions of the fault less commonly utilized in smaller ruptures as likely happened in Tohoku, the Haiyuan fault, and possibly in Cascadia. We suspect that elastic strain may accumulate at even very small rates in settings previously considered unfavorable, like a slowly charging battery, and that any such fault may be capable of eventually generating a very large earthquake using strain accumulated over many “normal” seismic cycles, and may then still have considerable energy remaining for more earthquakes. Such long-term cycling is suggested for the Himalayan front, the Haiyuan fault, NE Japan, and for Cascadia, one of a very few localities where enough events are recognized to test such models over many supercycles. A similar
relationship has been observed in Chile in the 1960 rupture zone (Cisternas et al., 2005), and supercycles may explain the periodic occurrence of the observed outsized events in Hokkaido.

At a minimum, the Tohoku earthquake implies that other comparable subduction zones, and perhaps others (McCaffrey et al., 2008) may be capable of similar behavior. Put in terms of relatively old subduction plate systems that were previously discounted as M9 producers, much of South America (possibly represented by the 1868 Arica earthquake, Chile; Dobath et al., 1990), the remainder of the Japan trench, the Kuriles, the western Aleutians, the Philippine, Manila and Sulu trenches, Java, the Makran and Hikurangi, Antilles and others may be capable of generating earthquakes much larger than known or expected today.

ACKNOWLEDGEMENTS

We thank Mary Lou Zoback for her thoughtful and detailed review that greatly improved the manuscript. Cascadia: The primary funding for field work and subsequent research has been provided by NSF Awards EAR 9803081, EAR-0001074, EAR-0107120, EAR-0440427 and OCE-0550843 (reservoir model development) and OCE 0850931 (2009 cruise). The U.S. Geological Survey substantially supported the work through Cooperative Agreements 6-7440-4790, 98HQAG2206, and 99HQAG0192; and the U.S. Geological Survey National Earthquake Hazard Reduction Program grants 02HQGR0019, 03HQGR0037, 06HQGR0149, and 07HQGR0064 to Goldfinger, and 02HQGR0043, 03HQGR0006, and 06HQGR0020 to Nelson. The American Chemical Society awarded support to Ph.D. student Joel Johnson for core collection and analysis at Hydrate Ridge under ACS PRF 37688-AC8.
REFERENCES

Atwater, B. F. (1987), Evidence for great Holocene earthquakes along the outer coast of

Atwater, B. F., and E. Hemphill-Haley (1997), Recurrence intervals for great earthquakes of the
past 3500 years at northeastern Willapa Bay, Washington, 108 p., U.S. Geological Survey,
Reston, VA.


Cisternas, M., et al. (2005), Predecessors of the giant 1960 Chile earthquake, *Nature*, **437**, 404-
407.

Clague, J. J., and P. T. Bobrowsky (1994), Tsunami deposits beneath tidal marshes on
Vancouver Island, British Columbia; with Suppl. Data 9421, *Geological Society of America

Chapman, J. S., and T. I. Melbourne (2009), Future Cascadia megathrust rupture delineated by


Goldfinger, C., et al. (2008), Late Holocene Rupture of the Northern San Andreas Fault and Possible Stress Linkage to the Cascadia Subduction Zone *Bulletin of the Seismological Society of America*, **98**, 861-889.


http://nisee.berkeley.edu/elibrary/Text/201204246


**Figure Captions**

Figure 1. (Left) Map showing recent vertical crustal movements and source areas of large interplate earthquakes. Blue line contours indicate rates of uplift (in mm/yr) revealed by tide gauge observations during the period 1955-1981 (Kato, 1983). Orange lines indicate source areas of interplate earthquakes of Mw > 7 since 1896. The epicenter and source area of the 2011 Tohoku earthquake of Mw 9.0 are indicated by an asterisk and orange shade, respectively NFM = Northern Fossa Magna. Open squares indicate tide-gauge stations; station numbers correspond to those in the right figure. (Right) Selected tide-gauge records along the Pacific coast (Geographical Information Authority, 2010). See the left figure for location. Red arrows indicate large earthquakes (Mw > 7.0) that occurred near each station. Note progressive subsidence of the Pacific coast at rates as high as 5-10 mm/yr, except for the Onahama station, which has likely been affected by coal mining.

Figure 2. Great earthquakes of the Himalayan front, showing meizoseismal zones of 1897, 1905, 1934, and 1950 earthquakes (ovals) and curved lines showing extent of larger earthquakes in 1100-1200 and 1505.
Figure 3. Correlation of four Cascadia turbidite cores spanning 550 km of the Cascadia margin from regionally correlated turbidites in Goldfinger et al. (2012). Events T11-T18 are shown to illustrate the two extreme events in this record, T11 and T16. T16 is a complex event with three elements at all sites, suggestive of shaking from three rupture patches in close succession. The three units in T16 are more widely spaced in proximal cores such as 56 PC. T16 also diminishes southward at Rogue Canyon. Location map show the four cores in large yellow-bordered symbols. Asterisk indicates benthic foraminiferal age corrected to planktic age.

Figure 4. Cascadia Holocene earthquake time series expressed as energy gain and loss per event. Energy gain is proportional to recurrence between events in years. Energy loss is proportional to the mass of turbidite samples, scaled to result in no net gain or loss of energy through the Holocene. Plot shows long-term energy cycling of the megathrust, and complex behavior over time. Mass values are shown dimensionless here, as they are extracted from the gamma density curves by digitizing the area under the mass curves for each event, using as a baseline the mass values for each baseline pair of bounding hemipelagic layers. Core compaction is partially compensated for by this method, though some compaction error remains in this plot. Error ranges are OxCal (Ramsey, 2001) 2σ ranges from Goldfinger et al. (2012) in the X axis (time), and an estimated maximum value of 10% error applied to the mass values that includes possible air gaps in the core liner, bioturbation if present within the turbidites (rare), measurement error, error in establishing the baseline for each turbidite, and digitizing error. Final event assumed to occur 500 years from last event at AD 1700, a value equal to the average recurrence time (Goldfinger et al., 2012).
Figure 1
Figure 2.
Figure 4