PALEOSEISMICITY OF THE CASCADE SUBDUCTION ZONE: EVIDENCE FROM TURBIDITES OFF THE OREGON-WASHINGTON MARGIN

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Abstract. Cores from Cascadia deep-sea channel contain sequences of turbidites that can be correlated and dated by the first occurrence of volcanic glass from the Mount Mazama eruption (6845 ± 50 radiocarbon yr BP). Turbidity currents from the tributaries appear to have occurred synchronously to form single deposits in the main channel, there being only 13 turbidite deposits in the lower main channel since the Mazama eruption, instead of the twice as many expected if the tributaries had behaved independently. In addition to the Cascadia Channel, 13 post-Mazama turbidites have been deposited in the Astoria Canyon and at two sites off Cape Blanco, sample locations that span 580 km of the Oregon-Washington margin. Pelagic intervals deposited between the turbidites suggest that in each place the turbidity currents occurred fairly regularly, every 500 ± 170 years on average. The best explanation of the spatial and temporal extent of the data is that the turbidity currents were triggered by 13 great earthquakes on the Cascadia subduction zone. The variability of turbidite timing is similar to that for great earthquake cycles. The thickness of the topmost pelagic layer suggests the last event was 300 ± 60 years ago (from three places along the margin), but this number may be a biased underestimate. It is, however, consistent with the youngest sudden-subside event on the Washington coast. The turbidite data demonstrate that the near-term hazard of a great earthquake on the Cascadia subduction zone is of the order of 2–10% in the next 50 years.

INTRODUCTION

Earthquakes have long been known to cause submarine disturbances sufficient to break submarine cables. The true nature of these events was not explained until the classic paper by Heezen and Ewing [1952] analyzed the sequential breakage of telegraph cables following the 1929 "Grand Banks" earthquake off Canada's eastern margin and showed that a submarine debris flow and subsequent turbidity current had travelled down the Laurentian Fan at speeds up to 65 km/hr and deposited a turbidite on the deep sea floor. Oceanographers in the 1950s and 1960s accord many turbidites in all the world's oceans, showed that they were chiefly associated with canyons and channels on submarine fans, and speculated on their rates of occurrence. Very high rates - every few years - off deltas such as the Magdalena and Congo suggested that a prime cause of the turbidity currents was sediment instability due to rapid sedimentation; the role of infrequent earthquakes was less easy to assess.

A classic study in turbidites and deep sea-sedimentation was carried out in the late 1950s by a group of students (Griggs, Nelson, Carlson, and Duncan) at Oregon State University under the supervision of L. D. Kulm. The most spectacular results came from the Cascadia Channel off the Oregon and southern Washington margin [Griggs, 1969; Griggs and Kulm, 1970]. Cores from the Cascadia Channel contained sequences of turbidites that could be correlated and dated by the first (lowest) occurrence of volcanic glass from the Mount Mazama eruption. The predominantly muddy sedimentation in the channel and the similarity of the cores suggest that they represent a complete record of turbidity currents in the Holocene. Here, 20 years later, the results are used to address the problem of great-earthquake hazard in the Pacific Northwest.
THE EARTHQUAKE PROBLEM IN THE PACIFIC NORTHWEST

About a dozen years ago, perceptions of the nature of the subduction zone beneath southern British Columbia, Washington and Oregon began to change [e.g., Riddihough and Hyndman, 1976]. Earlier work had established that the Juan de Fuca plate was converging toward North America. However the lack of seismicity on the plate interface lead some to consider that the subduction was occurring extremely slowly or had stopped.

More recent studies of the deformation front at the base of the slope [e.g., Barnard, 1978], of geodetic deformation rates on land [Ando and Balazs, 1979; Reilinger and Adams, 1982; Savage and Lisowski, 1988], and of onshore tectonic deformation such as warped terraces (see for example, Adams [1984], but also West and McCrum [1988a,b] and Atwater [1988a]) have confirmed that the Oregon-Washington margin is being deformed at rates as rapid as those other active subduction zones. The conundrum is this: with such high deformation rates, why have there been no large thrust earthquakes on the plate interface?

In a stimulating paper that addresses similarities between the Cascadia subduction zone and other zones worldwide, Heaton and Kanamori [1984] show that physical conditions imply that the subduction zone could generate earthquakes of magnitude 8.3 ± 0.5. Although the magnitude range they obtain is large, their study strongly suggests that infrequent great earthquakes are a distinct possibility, even though none have occurred historically.

Independently of Heaton and Kanamori, I suggested [Adams, 1984; 1985] that the simplest way to reconcile the discrepancies between long-term and short-term evidence for deformation, strain accumulation, and stress directions was if there was a great-earthquake cycle of long duration. I discussed how an analysis of landslides, landslide-dammed lakes, drowned trees below sea level, and uplifted beaches could be used to determine evidence for past great earthquakes, and mentioned the turbidites in the Cascadia Channel described by Griggs and Kulm [1970] as providing a minimum recurrence interval for such earthquakes.

The present paper analyses the turbidite record further, and shows that it provides good evidence for earthquakes about every 600 years. Such conclusions are important for the current debate [Heaton and Hartzell, 1987] on the level of seismic hazard in the Pacific Northwest, and are an important complement to the large body of onshore research into paleoseismicity that has occurred since 1980 [e.g., Atwater, 1987a,b].

SEDIMENT DISPERSAL OFF THE CASCADE MARGIN

To understand sedimentation off Oregon and Washington, it is necessary to understand the sediment sources, the processes by which sediment is carried to the deep-sea floor, and the temporal changes that have affected these processes. The chief source of sediment is the Columbia River, with lesser amounts of sediment coming from the Klamath Mountains in southern Oregon. Much sediment from the Oregon and Washington coastal ranges and from Vancouver Island is trapped in coastal inlets, and sediment from the Fraser River has been trapped in the Strait of Georgia for the last 8000 years.

At the present time, sediment from the Columbia River is carried north along the shelf and is deposited on the shelf and the edge of the slope (Figure 1). Sternberg [1986] reviews and cites the literature and infers that of the 21 x 10^6 t yr^-1 carried by the Columbia, 17% is deposited on the upper slope, most of it in the upper reaches of the Willapa, Grays, and Quinault canyons, which are tributaries of the Cascadia Channel, and some in the head of the Astoria Canyon. The substantial storage in the fluvial system and on the shelf is able to average out the short-term (10^-1 - 10^2 yr) climatic variations in supply.

The sediment accumulates on the slope until submarine failure occurs and the newly-accumulated sheet of sediment then sweeps down the channels as a density flow or turbidity current (Figure 2). Griggs and Kulm [1970] suggest that each major turbidity current takes about two days to travel the 735 km length of the Cascadia Channel and carries about 95 x 10^6 m^3 of sediment in a flow up to 17 km wide and 100 m high. Low deposition rates in the middle channel indicate that most of the sediment is carried as far as the Blanco Fracture zone, where it poudres, forming individual turbidite layers up to several meters thick [Griggs and Kulm, 1970].

The present dispersal pattern described above has lasted only for the past 6000-7000 years. Before about 7000 years ago, sea level was lower and still rising as water from the melting ice sheets was added to the oceans. During lower sea levels, the sediment was not carried north along the shelf to the same degree, but travelled directly to the deep sea down the Astoria canyon and channel. Cores on the middle Astoria Channel show little or no turbidite deposition in the last 7000 years [Nelson, 1976; although a core from the mouth of the Astoria Canyon (6500-PCI) shows many thin turbidites, presumably generated from the sediment spill-over from the Columbia shown on Figure 1. A typical turbidite in the Cascadia Channel system consists of an erosional base, a fine sandy basal layer and an upper layer of silt and clay that fines upwards. Because this sediment was derived from the biologically-rich outer shelf and upper slope, it contains shallow-water shell fragments and is high in carbon from partially decayed organic debris [Griggs et al., 1969, p. 168]. Once triggered, deposition from the turbidity current takes a few hours to days. Even the clay fraction must be deposited quite rapidly because deep-water currents would move any remaining suspended clays away from the newly-deposited turbidite. After all deposition has ceased, pelagic or hemipelagic sedimentation resumes, which is partly biological - foraminifera and radiolaria tests - together with small amounts of clay carried in suspension from the continent. In contrast to the turbidity currents which deposit 300 - 3000 mm in a matter of days, the pelagic sediment may accumulate at less than 0.1 mm yr^-1, slower farther away from the continent. Between turbidity currents there is colonisation of the newly-formed sea-bottom by burrow-
Fig. 1. Map of the Oregon-Washington margin showing the extent of the air-fall Mazama tephrula [Fryxell, 1965], the pattern of sediment dispersal north from the Columbia River (with amounts of sediment as percentage of the river input, after Sternberg [1986]), canyons on the Washington continental slope (J, Juan de Fuca, Q, Quinault, G, Grays, W, Willapa, and A, Astoria), submarine channels leading to the deep-sea floor, the location of cores mentioned (all from Oregon State University except core 53-18 which is University of Washington), and (within large circles) the number of post-Mazama turbidites as discussed in the text. The base of the continental slope (dashed line) and the 200-m isobath, marking the edge of the shelf, are also shown.
Adams: Paleoseismicity of the Cascadia Subduction Zone

Fig. 2. Perspective view showing schematically the genesis of turbidity currents in the Cascadia Channel. Sediment is carried down the Columbia River (1), drifts north along the shelf (2), and accumulates at the top of the continental slope in the heads of deep-sea canyons (3). According to the hypothesis presented in this paper, a great thrust earthquake on the subduction zone (4) causes strong shaking of the shelf and slope. The shaking causes sediment liquefaction and slumping simultaneously at many places along the margin. The resultant massive under-sea debris flows mix with the water to become a series of turbidity currents travelling synchronously down the channels (5). At junctions, the tributary turbidity currents coalesce to travel down the Cascadia Channel as one large turbidity current (6).

Fig. 3. Sketch of core 6609-24, made from the core log showing the even thicknesses of the turbidite and pelagic (shaded) layers. Only the top 6.4 m of the core is shown. The x’s in the 13th turbidite show the position of the earliest Mazama glass.

ing organisms (thriving on the high organic content of the turbidite silts) which cause bioturbation of the upper part of the turbidite and of the slowly accumulating pelagic sediment [Griggs et al., 1989].

The turbidites off Oregon and Washington have been sampled as 2 – 8 m long piston cores (e.g., Figure 3) taken in 2.0 – 2.5 km of water. For the purposes of this study the detailed core logs (which describe piston and associated pilot cores separately) compiled by students and staff at Oregon State University have been used to determine the turbidite and pelagic thicknesses shown on Table 1. Most of the cores still exist in storage at the university, but the logs have the advantage that they were written when the cores were freshly recovered, and were described by people with considerable experience. In almost all cases, the original colors and other evidence recorded in the logs are sufficient to determine the base and top of each turbidite, the thickness of the overlying pelagic sediment, and the amount of bioturbation; key parameters used in the present analysis.

CORRELATION AND DATING OF THE TURBIDITES BY THE MAZAMA TEPHRA

Because large turbidity currents travel down the length of the Cascadia Channel, each can leave a single deposit, and turbidite layers along the length of the channel should correlate (though some may be missing locally in parts of

the channel where erosion dominates). A unique event in the last 10,000 years was the eruption of Mount Mazama at 8845 ± 50 radiocarbon yrBP [Bacon, 1988]. During the brief eruption about 34 km³ of tephra ("volcanic ash") was deposited on Oregon and Washington. Sites on land show that prevailing westerly winds allowed little if any of the airfall tephra to fall as far west as the coastal ranges (Figure 1). Much, however, fell in the catchment of the Columbia River and this loose material was easily washed into the sea and deposited on the shelf. Experience with the much smaller Mount St. Helens eruption in 1980 showed that only about 16 months were needed for the tephra to travel to the head of Quinault Canyon [e.g., Sternberg, 1986], and there is no reason to think that the Mazama tephra would have behaved differently.

When the shelf-edge deposits slumped, the turbidity current carried the tephra-rich sediment down into channels. Cores taken along and near the channels contain the
### Table 1: Pelagic and Turbidite Thicknesses in Selected Oregon-Washington Cores

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Pelagic (left-hand) and turbidite (right-hand column of each pair) thicknesses (in mm) are given to the nearest 5 mm.

- 1° represents the topmost turbidite in each core.
- 2° Lowest post-Mazama turbidite (see text).

The tephras, but cores distant from channels do not [Nelson, 1976], confirming that air-fall tephras did not fall offshore and that deposition was via river, shelf, and channelled turbidity current.

By examining the coarse fraction at the base of each turbidite for fresh volcanic glass, it is possible to determine the first (lowest) turbidite that contains abundant fresh glass. Deeper layers contain sparse, weathered glass from eruptions older than the Mazama, but the first turbidite containing Mazama glass is very distinctive, and in some cases the sandy fraction is 25% glass (e.g., in core 6508-K1).

The first presence of the redeposited Mazama glass in the deep-sea turbidites provides time control with two important consequences: it allows the first turbidite with abundant glass to be traced down the Cascadia Channel, so confirming the nature of the event, and it allows calculation of the mean period between turbidity currents. Griggs and Kulm [1970] used the Mazama glass to compute recurrence intervals of 410 - 510 years for the turbidity currents in the Cascadia Channel.

In this paper I invert Griggs and Kulm's approach in order to discuss firstly the number of turbidity currents since the Mazama eruption, and then the mean recurrence interval. A typical core is shown in Figure 3. Turbidite and pelagic thicknesses and the lowest layer containing Mazama glass have been determined from the core logs, in almost all cases using the describer's boundaries (Table 1). In Cascadia Channel cores 6509-15 and 6609-24 the Mazama glass occurs in the 13th turbidite from the top; in cores 6705-6 and 6508-K1 it is in the 14th; for core 6705-5 there are no marginal notes on the core log indicating the position of the Mazama glass, but from the recurrence intervals given by Griggs and Kulm [1970, Table 3] it is in the 15th turbidite; and for core 6705-2 it is probably the 16th, but might be the 19th, as some of the sedimentary units are poorly defined.

The position of the Mazama glass in the three cores from the lower Cascadia Channel supports Griggs' [1969] assertion that not only must the lowest turbidite containing Mazama glass correlate, but so must each of the 12 overlying turbidites (i.e., the 13 turbidites in the cores represent the same 13 turbidity currents). Griggs' assertion is further supported by correlatable vertical variations in the degree of bioturbation in the top eight layers of cores 6509-27, 6609-24, and 6509-25 [Griggs et al., 1969]. Although the pattern of bioturbation is distinctive in these three cores and allows their correlation over 65 km of the lower channel, it is unlikely that similar correlations could be extended to the cores in the tributary channels (discussed be-
downstream is probably because one small turbidity current did not flow downstream to the confluence and to core 6508-K1.

The cores in the tributaries suggest that a few (2 or 3 out of a maximum of 16) small turbidity currents may have been generated that were smaller than the 13 turbidity currents recorded in the lower channel and did not travel as far. With this in mind, the remarkable inference can be made that pairs of turbidity currents were generated synchronously in the two tributaries. This arises because there have been 15 post-Mazama turbidites in the Willapa Channel and 14 turbidites in the northern channel, but only 14 turbidites below their confluence (where up to 29 might be expected if the flows had occurred independently). An alternative, though plausible explanation, is that exactly 15 of the turbidity currents, some from the Willapa Channel and the rest from the northern channel, died out before the confluence. This would require a higher rate of attrition than is evident in either the upper tributaries or in the lower Cascadia Channel.

Therefore turbidity currents were generated synchronously in two independent channels, the headwaters of which are a minimum of 50 km and a maximum of 150 km apart, and each pair of small turbidity currents merged to form one large turbidity current. That the number in the main channel is only 13 implies that every turbidity current in the two tributaries occurred synchronously, i.e. that synchronous occurrence is the rule rather than the exception. A similar argument can be made for other tributary channels that join the main channel farther downstream. Several of these tributaries contain 8-10 Holocene turbidites (e.g., cores 6609-27 and 6705-10), and yet the number of turbidites in the main channel does not increase beyond the 13 already present. Either all these turbidity currents predate the Mazama eruption or there were synchronous with the 13 turbidity currents in the main channel.

How closely in time did the tributary turbidity currents occur? If the turbidites were separated by 50 years or so, the pelagic drape (2 mm, at rates derived later in Figure 5) between the pair of events might be expected. For separations of a decade or so, there need be no perceptible pelagic layer, but some bioturbation in the lower turbidite could have occurred. For separations of a day or greater, each turbidite would still have left an individual fining-upwards sequence so each event would be a doublet: coarse base + fine top, coarse base + fine top, pelagic layer. Furthermore these doublets would persist down the Cascadia Channel. Examination of core 6508-K1 and the recent downstream does not reveal any double events, so I conclude that the tributary turbidity currents occurred less than a few days, and perhaps less than hours, apart.

My rough analysis of the time needed for turbidity currents to travel downstream from the canyon heads mapped on Figure 1 suggests that only a few tens of hours might separate currents from different, but synchronously-triggered sources at the canyon heads, so that the transport and deposition event in the main channel would be almost continuous. Using realistic estimates of flow velocity and sediment concentration, some tens of hours would be
needed to transport the sediment in an average Cascadia turbidite, which is consistent with with the above. Thus the coalescing flows combine to form the deposit described as a single turbidite. Similar coalescing of channelled turbidity currents to form a single turbidite deposit has been documented from the Mediterranean and from the Puerto Rico Trench [Pilkey, 1988], and has also been interpreted to represent the effects of a synchronous regional trigger.

**SPATIAL EXTENT OF THE 13 TURBIDITY CURRENTS SINCE THE MAZAMA ERUPTION**

Figure 1 shows three other cores that lie outside the Cascadia Channel system but contain the Mazama turbidite, one from Astoria Canyon, and two from the Blanco Valley physiographic province [Duncan, 1968] off Cape Blanco. In all three, the Mazama occurs in the 13th turbidite from the top.

The Astoria Canyon was a major conduit for sediment up until 7000 years ago. While Nelson [1976] shows that the lower Astoria Fan has been almost inactive since about the time of the Mazama (presumably due to the change in sediment dispersal that accompanied the sea level change at that time), core 6502-PC1 shows 13 thin (average thickness 130 mm) post-Mazama turbidity currents occurred in the upper channel. Apparently there is enough spill-over of the Columbia River sediment carried north along the shelf (Figure 1) for small turbidity currents to be generated, currents that do not travel all the way down the Astoria Channel. The thin turbidites are interbedded with unusually thick hemipelagic sediments, again due to closeness to the Columbia River source.

Both the core off Cape Blanco (6604-12) and the one off the Rogue River (6609-1) are remote from the Columbia River mouth, and heavy mineral analysis shows that at least since the Mazama eruption their sediment has come from the Klamath Mountains, inland and south of Cape Blanco [Duncan, 1968]. Despite their remoteness from the Cascadia and Astoria Channel systems they too contain 13 post-Mazama turbidites.

It is at least 50 km from the Juan de Fuca to the Quinault Canyon, 100 km across the Quinault-Willapa canyon system, 30 km more to the Astoria Canyon, 350 km farther to Cape Blanco, and then 50 km to the Rogue. From all five sites along the margin there have been 13 turbidity currents since the Mazama. I consider the simplest explanation for the synchronous events in neighbouring tributaries, and the same numbers of events at sites that span 550 km, is a series of great earthquakes.

**REGULAR RECURRENCE OF THE 13 TURBIDITY CURRENTS**

Up to this point the paper has discussed only the numbers of turbidites without regard to their timing. As noted by Griggs [1969] and Griggs and Kulm [1970], the turbidites in the Cascadia system have generally similar thicknesses and are separated by pelagic intervals also of generally similar thickness. The left half of Figure 4 plots turbidite and “pelagic” (hemipelagic plus pelagic) thick-
nesses for core 6609-24, in the lower Cascadia Channel. Both the turbidite (440 ± 130 mm each) and pelagic (55 ± 19 mm) layers are surprisingly regular in thickness (the standard deviation is 26% of the mean for the turbidite units and 35% of the mean for the pelagic units). For the pelagic thicknesses, the measurement error is about ±1/4 to ±1/3 of each value.

Some correlation between the thicknesses of adjacent pelagic and turbidite units might be expected. For example, an unusually long period between events would give both a thick pelagic unit (because of the longer time available for the pelagic sediment to accumulate on the deep sea floor) and a thick overlying turbidite unit (because of the longer time available for the slope sediments to accumulate in the canyon heads). However, this positive correlation is not evident on the upper right diagram of Figure 4. For some other cores (e.g., 6604-12) there is an inverse correlation between pelagic and overlying turbidite thicknesses, perhaps suggesting the thicker turbidites were deposited from larger turbidity currents that eroded more of the underlying pelagic sediments.

The variation of pelagic thicknesses for several cores can be seen conveniently by plotting cumulative pelagic thickness against turbidite sequence number (Figure 5). For this analysis to work, the rate of pelagic sediment accumulation should be steady, and the amounts of pelagic sediment eroded by subsequent turbidity currents should have been negligible or constant. The cumulative curves show similar linearity and scatter, supporting previous qualitative assertions that the events were fairly regular in time, and in particular that they did not cluster at the beginning or end of the sequence.

If turbidites in many cores were the result of single, spatially extensive events, there should be a correlation of deviations between the core plots, reflecting systematic differences in the intervals between the events. Correlations are not obvious, suggesting the signal/noise ratio is low, i.e., variability of timing (signal) is not very large relative to the imprecision in the measured pelagic thicknesses to estimate the intervals.

MEAN INTERVAL BETWEEN THE EVENTS AND VARIABILITY OF THE MEAN

The above argument suggests that 13 large events have shaken the margin. In addition to providing a correlatable horizon, the Mazama eruption is well-dated by multiple 14C dates at 645 ± 50 radiocarbon yrBP (Bacon [1983], though Brown et al. [1989] suggest that 640 ± 60 might be a better estimate), and provides a reliable date to compute the mean recurrence interval. The Bacon radiocarbon age is equivalent to 7000±150 (calibrated years before 1950) at the 2σ confidence level using version 2.0 of the computer program of Stuiver and Reimer [1983]). The difference between the upper and lower errors is small relative to the stochastic error discussed below, so adjusting for the 39 years since the 1950 reference date gives 7600 ± 100 calibrated years before 1989.

There is an unknown delay between the eruption and the deposition of the first turbidite containing Mazama glass. Because the St. Helens tephra was carried to the head of Quinault Canyon in 18 months, it is reasonable to suppose that Mazama tephra travelled similarly fast and was available for slumping almost immediately after eruption. When did the next turbidity current occur? For 13 turbidity currents the mean recurrence interval is about 600 years, so the next one would probably have been 500 ± 300 years later. Therefore, 7600 ± 400 years before 1989 is adopted as the age of the 13th turbidite.

One might speculate that the large volume of tephra must have caused more rapid accumulation on the shelf edge than normal, thus possibly reducing the time to the next sediment failure (which need not have been seismically induced). This might be the explanation for the '14th' turbidite in the upper Cascadia Channel (e.g., cores 6705-6 and 6508-K1). Furthermore, if the Mazama eruption were triggered by one great earthquake, the likelihood of the next would be reduced for about 600 years. Neither of these extreme adjustments to the age of the first post-Mazama turbidite can yet be proven.

Since the 13th turbidite there have been 12 turbidites. 12 inter-turbidite intervals and a period of time since the last turbidite. A priori, the length of the last period is not known so is taken as 1/2 ± 1/3 interval for a total of 12.5 ± 0.5 intervals. The mean interval between turbidites is then 7300 ± 400 years / 12.5 ± 0.5 intervals or 590 ± 50 years (for comparison, the Brown et al. [1989] age for the Mazama eruption gives 370 ± 60 years). Not used in this analysis is the information that if the last turbidity current was earthquake-triggered, it must have occurred more than 150 years ago, for there has been no great earthquake on the Cascadia subduction zone in historical times.

An independent check comes from core 6509-27 from which total carbon (mostly plant fibers) in the eighth turbidite from the surface was dated at 4645 ± 190 radiocarbon yrBP [Griggs et al., 1969], or about 5400 ± 500 calibrated years before 1989. The carbon had accumulated in the canyon heads in the 5000 years prior to the dated turbidity current, so applying a correction of ~300 years (half the inferred accumulation time) gives an interval of about 680 ± 150 years and by extrapolation beyond the base of the core, about 11 ± 2 turbidites since the Mazama. This is in good agreement with sedimentological correlations made by Griggs et al. [1969] on the basis of the degree of bioturbation, and the presence of 13 post-Mazama turbidites in the channel upstream and downstream of the core.

It is important to realize that the above error on the mean recurrence interval does not measure the variability of turbidity current timing. Such an estimate might be made directly by examining the thickness variations of the pelagic intervals. The thicknesses are not simple to determine because (1) the grain size and color of the uppermost turbidite layer and of the pelagic sediment is similar, and (2) bioturbation of the uppermost turbidite layer and the pelagic sediment confuses the lower boundary of the pelagic sediment. In addition, the present thickness may not record the original thickness because (3) the pelagic sediment may have been partially eroded during deposition of the overlying turbidite, and (4) bottom cur-
Fig. 5. Cumulative pelagic sediment deposition versus turbidite number for seven cores from the Oregon-Washington margin. The plots are lined-up on the turbidite containing the first Mazama glass (star), and penultimate and top pelagic layers are labelled by P and T respectively. If the pelagic sediment has accumulated steadily, and if the turbidity currents occurred regularly, each plot would be a straight line. The data (dots joined by heavy lines) shows a good approximation to the expected relationship (light lines). Note that the vertical scale has been adjusted by factors of 2 for three cores so that all the curves have a similar slope. The slope of the fitted line represents the mean pelagic sediment accumulation rate, and the deviations of the data about the line gives information on the variability of turbidity current timing, here expressed as a percentage of the mean pelagic accumulation per event.
rents may have eroded or winnowed the pelagic layer as it accumulated [Harllett and Kulm, 1972]. All of these factors would tend to reduce the thickness of the pelagic layer, but are difficult to quantify unless good estimates of the pelagic accumulation rate can be applied. The rather constant thickness of the pelagic layers suggests that the losses, if any, are rather constant. In any event, the variance from the four sources above must be added to the variance due to irregular event occurrence, so that the events must have been more regular than is implied by the variability of the pelagic thicknesses.

For example, the numbers against the plot for each core on Figure 5 represent the best-estimate mean pelagic accumulation per event and the standard deviation expressed as a percentage of the mean. Both accumulation rates and their variability are generally largest near shore, while the deeper cores are the most regular, ±26% for 6508-K1 and ±36% for 6509-15 (not shown on Figure 5). If the cores are considered to represent the same 13 events, those with the more variable pelagic thicknesses must include a larger amount of variance due to measurement errors and erosion, as discussed above, and so provide poor estimates of the variability of event timing. Therefore, the average variability for the three lowest cores (1σ = 26%) is chosen as representative. For a 930-year recurrence interval, this translates to a variability (1σ) of about 170 years, and an implied standard error on the mean of 50 years. Therefore, if the intervals are normally distributed there is one chance in two that one of the 12 intervals would lie outside $\mu \pm 1.7\sigma$, or outside the range 300 to 900 years.

DO THE TURBIDITES REPRESENT GREAT EARTHQUAKES, OR NOT?

Although G. Griggs originally believed that earthquakes were the cause of the turbidity currents (L. D. Kulm, personal communication, 1988), the lack of large earthquakes nearby made this position difficult to support in the middle-to-late 1960s. Griggs and Kulm [1970] noted that if sediment is supplied by the Columbia River at a constant rate, it would accumulate in the canyon heads until there is enough to trigger collapse, whereupon the cycle would start again. Such a self-triggering system would tend to repeat itself, and would generate similar-sized turbidites in the channel system; to borrow a term from seismology, it would generate the “characteristic” turbidite in the channel system.

Of course, triggering by external events of a cyclic nature (such as great earthquakes) would also tend to displace similar masses of sediment and produce similar-sized turbidites. Perhaps the strongest argument against the self-triggering hypothesis for the Cascadia margin is that similar numbers of events are found at widely separated places along the margin; it would be highly improbable that every canyon had the same temporal response to the spatially-varying rates of sedimentation that occur along the margin. It is simpler to conclude that, while self-triggering might occur in the absence of an external trigger, the triggers occur more frequently than the time needed to reach the critical mass in most canyons, and so usually short-circuit the endogenic slumping process.

For a similar reason, events local to one canyon head (e.g., a magnitude 6.5 earthquake within 50 km) do not provide a sufficient explanation for synchronous turbidity currents in separated tributaries and the same numbers of events along the margin. Therefore the cause must be exogenic to the canyon heads and affect the Oregon-Washington margin as a whole. Three such external triggers with the required spatial extent are: tsunamis, wave-induced slumping during large storms, and great earthquakes on the Cascadia subduction zone.

Tsunami. The most recent damaging tsunami on the Oregon-Washington coast was generated by the 1964 Alaska earthquake, the second largest earthquake of this century. It did not trigger one of the 13 great turbidites in the Cascadia Channel, because both the amount of pelagic sediment on top of the cores (see below) and the amount of bottom life on the channel floor during the 1965-1967 cruises [e.g., Griggs et al., 1969] are too large for such a recent turbidity current. Tsunami-generating earthquakes are relatively common around the Pacific, and if the large tsunami in 1964 did not trigger a turbidity current, it seems unlikely that a coupling of multiple independent tsunami sources with the time-dependent stability changes in the canyon heads would give both long recurrence intervals and the same number of events along the margin.

Wave-induced slumping. The case for wave-induced slumping on such a large scale is poorly documented, and computations suggest that for all but the steepest slopes wave loading does not influence slope stability in water depths greater than 120 m [Moran and Huribut, 1986], i.e., shallower than most of the canyon heads. Physical parameters constrain the maximum size of storms and storm-generated waves so that the rate of size increase falls off rapidly with decreasing probability (e.g., the 1000-year wave may be only 30% larger than the 100-year wave, and not ten times the size). Therefore the 100- and 1000-year storms would have rather similar effects, it is likely that the sediments would need to be close to failure anyway, and this would happen at different times for different places. Like tsunamis, each large storm would trigger some turbidity currents when individual canyon heads were ready, and so the process would be unlikely to give both long recurrence intervals and the same number of events for the whole margin.

Earthquakes. Earthquakes are an unusual natural phenomenon in that even for quite rare events the expected ground shaking increases dramatically as the probability level drops. Thus a great earthquake is so overwhelmingly large that it would trigger both marginally stable and “stable” canyon head deposits. Great earthquakes, should they occur on the Cascadia subduction zone, would probably have a long return period, and indeed return periods of several hundred to a thousand years have been estimated by Adams [1984] and Heaton and Hartzell [1986], based on the rate of geotectonic strain accumulation. As a first approximation, plate-boundary earthquakes occur fairly regularly (a consequence of the steady build-up of strain due to plate movement and the constant physical parameters of the fault zone), and so tend to have a “characteristic” size. Very short intervals are unlikely because insufficient strain is available to be released, and very long intervals
are unlikely because the strength of the fault zone limits the amount of strain that can be stored. For such characteristic earthquakes there may be a scale-independent variability in timing amounting to about ±20% of the mean recurrence interval [Nishenko and Buland, 1987], 120 years for a 590-year recurrence interval. This is close to the 20-
30% variability in pelagic thicknesses found in the cores (which includes not only variability due to the earthquake cycle but also that due to variable partial erosion and measurement errors in determining the thickness of the pelagic layers).

Evidence that the 13 turbidity currents indeed represent earthquakes is circumstantial and proof is unlikely for several years. Nevertheless, because a great thrust earthquake is an extremely large and relatively rare event, it should cause synchronous effects throughout the coastal Pacific Northwest. Such confirmatory effects would include: sudden coastal subsidence or uplift, landslides, sediment liquefaction and sand volcanoes, sediment slumping in large lakes, and submarine debris flows. Other indirect evidence would be: abnormal sedimentation events, deformed tree growth, abandonment of Indian settlements, and secondary faulting on crustal faults. It should also be noted that while the turbidite record may well be complete because of the muddy nature of the sediments, the inferred earthquake record might be incomplete because subsequent near-by earthquakes (a second mainshock or large aftershock) soon after a great earthquake would probably not generate their own turbidity current; all of the unstable sediment having already slumped.

The first onshore evidence inferred to indicate past great Cascadia earthquakes was reported by Atwater [1987a] who ascribed buried Holocene marshes in coastal Washington to sudden coseismic subsidence of the coast. His most recent work [Atwater, 1988a] suggests five subsidence events and one shaking event in the last 3100 years. Corresponding evidence comes also from eight buried soils in about 5000 years from another coastal Washington site [Hull, 1987], seven buried marshes in about 3500 years from northern Oregon [Peterson et al., 1988], and eight buried marshes in about 5000 years from southern Oregon [Nelson et al., 1989]. At the southern end of the Cascadia subduction zone, Carver and Burke [1987] deduced recurrence intervals of about 600 years for thrust faults in the accretionary prism just north of Cape Mendocino. In each place the implied recurrence interval is about 500 – 600 years, in good agreement with the mean recurrence interval and spatial extent established for the turbidity currents (Figure 6a).

B. Atwater [personal communication, 1988] has also drawn my attention to the coincidence between the age of 4290 ± 80 radiocarbon yrBP for the 8th peat at Willapa Bay (sample H1-B of Atwater [1988b]) and the age of 4645 ± 100 radiocarbon yrBP from the 8th turbidite from core 6502-27 (discussed previously). Applying the correction of -300 years to the core date brings the two ages into remarkable agreement. This agreement is all the more significant because each date is on the eighth event – thus suggesting a one-to-one correspondence between the turbidite and subsidence events, and reinforcing the argument made from the mean recurrence intervals above.

Fig. 6. Spatial distribution of derived values for (a) Recurrence intervals for turbidites and onshore subsidence events, and (b) Age of the last turbidite, submarine slumping, or onshore subsidence event. Data taken from sources discussed in the text.

While I do not claim that there is yet any completely convincing proof that the 13 turbidity currents do represent 13 great thrust earthquakes on the Cascadia subduction zone, I believe that it is the best interpretation available at present.

IMPLICATIONS FOR GREAT THRUST EARTHQUAKES ON THE CASCADE SUBDUCTION ZONE

The occurrence of 13 great thrust earthquakes, their mean recurrence interval of 590 years, a variability of 170 years, and the assumed 580 km extent of triggered turbidity currents and hence inferred strong ground motion that are deduced above place some constraints on the slip, the rupture dimensions, magnitude, and style of rupture of the subduction zone. Rogers [1988a] has divided the Cascadia subduction zone into plausible segments. The Juan de Fuca plate extends 900 km from the Nootka Fault Zone off Vancouver Island to a boundary with the South Gorda plate off northern California. An alternative segmentation stops the Juan de Fuca plate at the Blanco Fracture Zone.

Subduction of the Juan de Fuca plate at 45 mm/yr and the mean recurrence interval of 590 years gives 26 m for the average slip per earthquake, provided no slip is aseismic. A 20 m slip for the 750 km long Juan de Fuca plate north of the Blanco Fracture Zone and a fault width of 100 km [Rogers, 1988] gives a maximum moment magnitude (Mw) of 9.1, which is in accord with the maximum size proposed by Rogers [1988]. A variability of 25% in event timing would imply slip displacements of between 19 and 33 m, but a variability in magnitude of only ±0.1 magnitude unit.

Would a rupture of the Juan de Fuca subduction zone stop at the Blanco Fracture Zone? Key evidence comes from core 6604-12, which lies on the extension of the frac-
ture zone. This core - as does core 6609-1 to the south, on the subducting Gorda plate - contains 13 post-Mazama turbidites, here presumed to represent great earthquakes on the Juan de Fuca subduction zone. If the Gorda plate had a history of independent subduction (e.g., in Mw 8.3 earthquakes; Rogers [1988]) core 6604-12 is close enough that additional turbidity currents might have been generated by earthquakes to both north and south. Therefore the presence of only 13 events in these two cores suggests that either every Juan de Fuca rupture extends past the Blanco Fracture Zone (i.e., a Mw 9.2 earthquake), or the earthquakes on the Gorda segment are synchronized with those on the Juan de Fuca (i.e., zipper effect in which an earthquake on one segment triggers earthquakes on adjacent segments in succession for a few hours to years later). In the latter case insufficient sediment would have accumulated between the paired earthquakes for a turbidity current to have been generated by each earthquake.

WHEN WAS THE LAST EVENT?

The age of the last event can be estimated roughly from the thickness of the topmost pelagic sediment in the cores relative to the mean accumulation rate. As noted by Griggs and Kulm [1970, p. 1387] "A lack of recognizable hemipelagic sediment at the surface of these cores and the absence of extensive burrowing in the surficial layer compared to the extensive reworking in lower layers indicate that the last flow was recent", although the fact that all their cores have thinner last turbidites than penultimate turbidites suggests that the very soft pelagic sediment and the top, bioturbated part of the most recent turbidite were often washed away. In addition, their use of "recent" is in the context of Holocene time, and might be taken to include the last 1000 years.

The thickness of the topmost pelagic sediment, from the available core logs which include "Phi" trigger-weight cores taken with the piston cores, has been used together with the long-term pelagic accumulation rates (some shown on Figure 5) to compute the time since the last turbidite. From six cores (Table 2) an age of about 300 ± 60 years (before 1989) is estimated. This is consistent with the lack of a great earthquake in the 150-year historical record, but may be a biased underestimate if some sediment washed out from the top of each core. It is noteworthy that both the Astoria and the Blanco cores give similar (to within the poor resolution) ages to the four Cascadia cores (Figure 6b), thereby showing that they did not occur at greatly different times (though this is not proof that they occurred at the same time), and perhaps suggesting a regional correlation of the last event.

A date for the last turbidite could also be derived from the amount of sediment that has accumulated in the submarine canyons since the sediment last slumped away. In Quinault Canyon piston core 68-18 only 200 mm of hemipelagic sediment has been deposited since the last flushing of the canyon [Barnard, 1978, p. 111], representing 400 years (using Barnard's rates) or 100 years of sediment accumulation (using rates of Thorbjarnarson et al. [1986]). Cores 53-19 and 63-08 contain 200 mm and 350 mm of post-flushing sediment [Barnard, 1978], representing 200-400 years accumulation (using rates of Thorbjarnarson et al. [1986]). While improving the precision of these estimates clearly requires a critical assessment of the sedimentation rates at the core sites in question, the above rates generally support the inference that 300 years or so have elapsed since the last event.

Tree-ring dates on drowned trees at six sites in coastal Washington suggest the last synchronous, rapid submergence event was about 300 years ago [Yamaguchi et al., 1989]. Comparable dates on the youngest buried marsh soil suggest the last subsidence event occurred about 300 years ago in southwest Washington [Atwater et al., 1987], less than 400 ± 200 years ago in northern Oregon [Peterson et al., 1988], less than 380 ± 60 years ago in southern Oregon [Nelson et al., 1988], and 270 ± 30 years ago in northern California (G. Carver, personal communication, 1989); dates that are consistent with the preceding analysis (Figure 6b).

TABLE 2. Age of the Last Turbidite From Thickness of the Topmost Pelagic Sediment

<table>
<thead>
<tr>
<th>Core #</th>
<th>Pelagic Thicknessa (mm)</th>
<th>Accumulation Rateb (mm/event)</th>
<th>Age of Last Turbidite (years)c</th>
</tr>
</thead>
<tbody>
<tr>
<td>6705-2</td>
<td>23 ± 6</td>
<td>62</td>
<td>220 ± 60</td>
</tr>
<tr>
<td>6705-6</td>
<td>25 ± 3</td>
<td>55</td>
<td>270 ± 30</td>
</tr>
<tr>
<td>6509-26</td>
<td>17 ± 5</td>
<td>41</td>
<td>245 ± 70</td>
</tr>
<tr>
<td>6509-27</td>
<td>16 ± 3</td>
<td>28</td>
<td>335 ± 60</td>
</tr>
<tr>
<td>6502-PC1</td>
<td>70 ± 20</td>
<td>148</td>
<td>280 ± 80</td>
</tr>
<tr>
<td>6604-12</td>
<td>47 ± 10</td>
<td>105</td>
<td>260 ± 60</td>
</tr>
<tr>
<td>mean</td>
<td></td>
<td></td>
<td>270 ± 60</td>
</tr>
</tbody>
</table>

---

aThickness (± possible error) in 1965/1967.

bFrom Table 1 and Figure 5 and similar plots.


dOr about 300 years before 1989. May be biased toward being too young because of washouts at top of core; see text.
TABLE 3. Age of the Last Five Events in Core 6508-K1

<table>
<thead>
<tr>
<th>Pelagic (mm)</th>
<th>Turbidite (mm)</th>
<th>Δt* (years)</th>
<th>Ageb (years)</th>
<th>Date</th>
</tr>
</thead>
<tbody>
<tr>
<td>10</td>
<td>160+</td>
<td>300±20°</td>
<td>300±60</td>
<td>A.D. 1690±60</td>
</tr>
<tr>
<td>22±3</td>
<td>240</td>
<td>300±70</td>
<td>690±130</td>
<td>A.D. 1300±130</td>
</tr>
<tr>
<td>28±3</td>
<td>570</td>
<td>500±70</td>
<td>1190±200</td>
<td>A.D. 800±200</td>
</tr>
<tr>
<td>55±12</td>
<td>120</td>
<td>980±250</td>
<td>2170±450</td>
<td>180±450 B.C.</td>
</tr>
<tr>
<td>34±3</td>
<td>65</td>
<td>610±70</td>
<td>2280±520</td>
<td>790±520 B.C.</td>
</tr>
</tbody>
</table>

*Time interval represented at 33±1 mm/590-year event assuming no loss of sediment.
*bCalendar years before 1989.
See Table 2.

Working from the pelagic thicknesses in core 6508-K1 (chosen because its pelagic thicknesses are the most regular), assuming that no sediment has been eroded by the overlying turbidite, and adopting 300 years for the age of the last event, the following dates for the last five earthquakes are deduced (Table 3): 1690 AD, 1300 AD, 800 AD, 180 BC, and 790 BC. Although the slow accumulation rates of the pelagic sediment and the disturbance by the corer mean these dates are not very precise (and in particular, the errors compound for the older dates as shown in Table 3), they should prove of interest to other researchers seeking to match their onshore events to the turbidite record.

The analysis presented here underlines the need to collect new cores from the Cascadia Channel and the remainder of the margin. A profile of closely-spaced cores across the channel, somewhere around core 6508-K1 on Figure 1 would reveal much about the history of the turbidity currents. Griggs and Kulm [1970] obtained several cores from the levees of Cascadia Channel banks that contained only 8 post-Mazama turbidites; presumably the largest ones. At some point between the levees and the channel floor, there should be a complete section of the 13 turbidites, from which each subsequent turbidity current has eroded little or none of the pelagic sediment. Such cores would be invaluable for establishing the relative sizes and timing of the 13 turbidity currents. It might also test whether the amount of bioturbation is an indicator of elapsed time or relates to the size of the previous turbidity current. In addition, radiocarbon dating of detrital carbon or even of individual pelagic foraminifera in the cores would place independent ages on the events and avoid the compound errors intrinsic in an analysis like that in Table 3. In the upper channel and on the slump scars of the canyon heads, box cores, or other non-disturbing sampling methods, should be used to determine the thickness of the topmost sediment in order to establish the time since the last event.

PROBABILITY OF THE NEXT EVENT AND SEISMIC HAZARD IMPLICATIONS

From the mean interval between events (590 years), the standard deviation of the mean (±170 years), the time since the last event (±300 years), and a normal-distribution model for simple recurrent ruptures without clustering, it is possible to estimate the likelihood of the next great Cascadia earthquake (Figure 7). At present there is about a 5% chance that the next earthquake should have already happened. For the future, the conditional probabilities are crudely 0.1% in the next year, 5% in the next 50 years, 10% in the next 100 years, and 25% in the next 200 years.

These values are probably good to only a factor of 2 and

![Fig. 7. Graph showing the cumulative normal probability distribution for a mean of 590 years and a standard deviation of 170 years as inferred for great earthquakes on the Cascadia subduction zone, an estimate of where the present lies relative to the last event (thick bar on the abscissa at 300±60 years), and the range of conditional probabilities of the next event for the next 50, 100, and 200 years.](image-url)
need to be refined using a revised date for the last earthquake and perhaps a lognormal distribution model (following Nishenko and Buland, 1987); however some of the uncertainty arises from the possibility that the events cluster in time (say three events in one millennium followed by a long quiescence) and from the still-unknown mode of failure of the subduction zone. While the same number of turbidites in the Cascadia Channel, Astoria Channel, and the sites off Cape Blanco strongly suggests synchronous turbidity currents, it is not yet possible to rule out a zipper effect whereby smaller earthquakes rupture the plate boundary in a short-term sequence. Such multiple modes of rupture—sometimes single great earthquakes, sometimes sequences of smaller earthquakes—would complicate the paleoseismic history and are probably beyond the resolving power of the turbidite record, though could be resolved through onshore investigations using dendrochronology or other precise dating of earthquake effects.

Even if the rupture mode is complex, the net effect on hazard estimates may not be great because circumstantial evidence suggests that the (hypothetical) rupture segments would need to ‘stay in step’ and the time since the last event is already close to half the mean recurrence interval. In addition, the damage implications for any place along the margin are not greatly different for the single M9 or multiple M8 earthquake scenarios; though clearly one M9 earthquake happen somewhere on the subduction zone the likelihood of others would be increased because of the temporal clustering implied by the alternate hypothesis.

CONCLUSIONS

Turbidites in the tributaries of the Cascadia Channel and at other places along the Oregon-Washington margin provide circumstantial evidence for the occurrence of 13 great Cascadia subduction zone earthquakes since the Mazama eruption. My analysis suggests that magnitude M9 earthquakes occurred every 500 years on average. The pelagic intervals deposited between the turbidites suggest that the earthquakes occurred fairly regularly, with a standard deviation of 170 years or less on the recurrence interval, similar to the variability found for great earthquake cycles elsewhere.

Rhythmic triggering of turbidity currents by great earthquakes may be a much more common phenomenon than hitherto realised, and might be expected at continental margins such as Alaska, Japan, New Zealand, and Chile where great thrust earthquakes with a long return period are combined with a moderate supply of sediment to the edge of the shelf. If sampled correctly, the turbidite record can provide a quick estimate of the paleoseismicity of a margin and so provide evidence independent of onshore paleoseismicity studies.

The thickness of the topmost pelagic layer suggests the last earthquake was 300 ± 50 years ago, but this number may be a biased underestimate due to washout at the top of the core during the collecting process. It is, however, consistent with the youngest subsidence episode on the southwest Washington coast. The pelagic accumulation between the turbidites has the potential to determine the timing of previous events by working backwards; however, in the absence of absolute dates on the turbidite layers in the cores (like those available from the submerged peats onshore), the errors compound and such dates should be used with caution. The current state of great earthquake expectation for the Pacific Northwest (Figure 7) demonstrates that the near-term hazard of a great earthquake is appreciable.

Material capable of precisely dating the past earthquakes has already begun to be studied at onshore sites where coseismic deformation and tsunamis can be dated by radiocarbon or dendrochronology. With turbidites to provide an independent record of shaking, the concordance of diverse data may ultimately reveal the history of great thrust earthquakes in the Pacific Northwest.

Acknowledgments. This work would not have been possible without the meticulous work of G. B. Griggs, his contemporaries, and their supervisor La Verne Kulm. Their work, which I have reinterpreted in this paper, is proof that good science has ramifications well beyond the aspirations of the individual scientist. I thank Mitch Lyle and the College of Oceanography for their cooperation and for making the core logs available. I apologize to my colleagues who have heard me present this work at the 1984 Chapman Conference on Vertical Crustal Motion and at subsequent meetings, who have only now a written version to examine critically, but who have nevertheless carefully cited and attributed my work. D. Piper assisted with technical advice regarding the behaviour of turbidity currents and B. Atwater pointed out the importance of the bias in the uncalibrated radiocarbon ages. I thank P. W. Basham, M. J. Berry, D. J. W. Piper, and G. C. Rogers for valuable critical comments on early versions of this manuscript, and B. Atwater and L. D. Kulm for constructive reviews. Geological Survey of Canada contribution number 33288.

REFERENCES

Adams: Paleoseismicity of the Cascadia Subduction Zone


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(Received September 30, 1988; revised December 12, 1988; accepted December 18, 1988.)