Variation of Modern Turbidite Systems Along the Subduction Zone Margin of Cascadia Basin and Implications for Turbidite Reservoir Beds

C. Hans Nelson
U.S. Geological Survey and Department of Oceanography
Texas A & M University
College Station, Texas 77843
409 458 1816
e-mail: hans@ocean.tamu.edu

Chris Goldfinger
College of Oceanic and Atmospheric Sciences
Oregon State University
Corvallis, Oregon 97331

Joel E. Johnson
Department of Geosciences
Oregon State University
Corvallis, Oregon 97331

Gita Dunhill
INSTARR/ Department of Geological Sciences
University of Colorado
Boulder, Colorado 80309

Abstract

Cascadia Basin contains a variety of turbidite systems located from Vancouver Island, Canada to Cape Mendocino, California, USA. These systems have been studied with multibeam bathymetry, sidescan sonar, high-resolution seismic profiles, and piston cores. On the Washington margin, multiple canyon sources funnel turbidites into Cascadia Channel, a single high-relief deep-sea channel, that extends across Cascadia Basin and cuts through Blanco Fracture Zone. Astoria Canyon feeds Astoria Fan, a submarine fan with channel splays and depositional lobes which fill the subduction zone trench off Oregon. Both of these large turbidite systems (1000 km length) prograde mainly southward parallel to the margin in northern Cascadia Basin. In south Cascadia Basin, small turbidite systems (5-50 km) prograde perpendicular to the margin. Rogue Canyon feeds a small (~5 km) base-of-slope apron. Trinidad and Eel canyons feed into plunge pools and sediment wave fields that extend tens of km radially out from the canyon mouth. A channel-levee complex drains the Eel sediment waves and feeds into a sand-rich lobe. Mendicino Channel, a connecting channel-levee complex without distal lobes, traverses the base of Mendicino Escarpment at the triple junction. Turbidite systems from the Rogue River north contain 13 correlative post-Mazama turbidite events based on the first occurrence of Mazama Ash (MA) at about 7530 calendar yr BP. Another 12,300 calendar yr datum, at approximately the Pleistocene/Holocene boundary (H/P), is found throughout Cascadia Basin. Based on these datums, turbidite events appear to be triggered by seismic events on average every 600 years in northern Cascadia Basin and progressively more often toward the Mendocino Triple Junction (i.e in Trinidad pool every 492 yr, in Eel lobe every 246 yr and in Mendocino Channel every 40-65 yr).

The correlation of turbidite events can be used to compare bedding continuity within systems and between different systems to provide important implications for turbidite reservoir characteristics. The progressive loss of post MA turbidites down the proximal 150 km of Astoria Channel suggests that during this time, downfan continuity in turbidite beds is less in fan channels compared to Cascadia Channel where all 13 post-MA beds are continuous throughout the deep-sea channel. In contrast, both deep-sea and fan channels exhibit cut and fill in proximal regions, sediment bypassing and down channel dropout of beds during the Pleistocene. As a result, high sand-shale...
rations (1:1 to 3:1) are found in distal fan lobes during the Pleistocene whereas low rations are found during the Holocene. Good lateral bedding continuity is found throughout the Rogue apron that is undisturbed by channels. Turbidite events are twice as common in plunge pools compared to the downstream sediment waves, which suggests a loss of bedding continuity in sediment waves that is analogous to that in channel levees. However, in the case of the Eel system, when the pool and waves are drained by a channel-levee complex, the highest frequency of turbidite beds and sand-shale ratios (1:8:1) are found in the distal lobe. Sand-Shale ratios and frequency of events suggest that during the Pleistocene, sediment erosion and bypassing took place in the pools compared to the infilling of the Holocene. The greatest Holocene infilling rate takes place in Mendocino Channel where turbidite events occur every few decades and sand-shale ratios are 2.5:1.

**Introduction**

Cascadia Basin covers the deep ocean floor over the Juan de Fuca and Gorda Plates, and extends from Vancouver Island, Canada to the Mendocino Escarpment in northern California, USA (Fig. 1). Cascadia Basin contains a wide variety of Quaternary turbidite systems that exhibit different patterns of channel development and turbidite sand-bed distribution (Fig. 2). Study of these systems with multibeam bathymetry, sidescan sonar, high-resolution seismic profiles, and accurately sited piston cores show that there is considerable variation in distribution patterns of sand-prone reservoir facies for the different types of turbidite systems. These varying turbidite system architectures have been deposited along the 1000 kilometer Cascadia margin that has the same tectonic and environmental setting. This variation of systems along the Cascadia margin needs to be considered when developing reservoir models for turbidite system analogues in the Cascadia margin. This terrane may have been accreted to the margin because of this unique correlation. Our final synthesis of the turbidite systems in Cascadia Basin combines these new data, particularly from the small sand-rich Rogue, Trinidad, Eel and Mendocino turbidite systems, with the extensive studies that have previously been published on the Cascadia, Nitinat, and Astoria systems (Nelson et al., 1968, 1970, 1987; Duncan, 1968, Griggs and Kulm, 1970; Carson, 1973; Nelson, 1976).

**Subduction Zone Setting and Earthquake Potential of Cascadia Basin**

The Cascadia subduction zone is formed by the subduction of the Juan de Fuca and Gorda oceanic plates beneath the North American plate off the coast of northern California, Oregon, Washington, and Vancouver Island (Fig. 1). The convergence rate is 40 mm/yr directed N68°E at the latitude of Seattle (DeMets et al., 1990). The submarine forearc widens from 60 km off southern Oregon to 150 km off the northern Olympic Peninsula of Washington, where the thick Pleistocene Astoria and Nitinat Fans are presently being accreted to the margin. Off Washington and northern Oregon, the broad accretionary prism is characterized by low wedge taper, and widely spaced landward vergent accretionary thrusts and folds which escarp virtually all of the incoming sedimentary section. Sparse age data suggest that this prism is Quaternary in age, and is building westward at a rate close to the orthogonal component of plate convergence (Westbrook, 1994, Goldfinger et al., 1996). This young wedge abuts a steep slope break that separates it from the continental shelf. Much of onshore western Oregon, Washington and the continental shelf of Oregon is underlain by a basement of Paleocene to middle Eocene oceanic basalt with interbedded sediments, known as the Crescent or Siletzia terrane. This terrane may have been accreted to the margin (Duncan, 1982), or formed by in situ rifting and extension parallel to the margin (e.g. Wells et al., 1984). Much of the Oregon and Washington shelf is underlain by a modestly deformed Eocene through Holocene forearc basin sequence.

The continental slope of Cascadia Basin is traversed by numerous submarine canyons delivering an abundant sediment supply to the filled trench from the high-rainfall coastal region and continental interior (Fig. 1). The Columbia River, one of the largest rivers in North America, has delivered the greatest sediment load of about 20 million metric tons per year during late Holocene (Wolff et al., 1999). During the Pleistocene low-stands, when there was glaciation of much of the Columbia River drainage, a greater sediment load was delivered directly to the Cascadia Basin floor turbidite systems, constructing systems.
Figure 1. Physiographic map showing the plate-tectonic setting of Cascadia Basin and the 1999 core locations. Northern Cascadia Basin overlies the part of Juan de Fuca Plate that extends from Juan de Fuca Ridge and Blanco Fracture Zone to the deformation front of Cascadia Subduction Zone at the base of the continental slope off Vancouver Island, Washington and Oregon. Southern Cascadia Basin overlies the part of Gorda Plate that extends from Gorda Ridge and the Mendocino Escarpment to the deformation front of Cascadia Subduction Zone at the base of the continental slope off southern Oregon and northern California.
Figure 2. Map showing location of the different types of turbidite systems within Cascadia Basin. Locations are indicated for selected cores that are exhibited in Figures 4, 5, 6, 12 and 14. For location of all general seafloor morphologic features and contour depths see Figure 1.
such as the broad Astoria Fan that filled the subduction zone trench (Nelson et al., 1987). During the Holocene, sediments from Cascadia rivers have been deposited in nearly full shelf basins and upper canyons (Goldfinger et al., 1992).

The earthquake potential of Cascadia Basin margin has been the subject of major paradigm changes in recent years. First thought to be aseismic due to the lack of historic seismicity, great thickness of subducted sediments, and low uplift rates of marine terraces (Ando and Balazs, 1979; West and McCrumb, 1988), Cascadia is now thought capable of producing large subduction earthquakes on the basis of paleoseismic evidence (e.g. Atwater, 1987; Atwater et al., 1995; Darienzo and Peterson, 1990; Nelson et al., 1995), geodetic evidence of elastic strain accumulation (e.g. Mitchell et al., 1994; Savage and Lisowski, 1991; Hyndman and Wang, 1995) and comparisons with other subduction zones (e.g. Atwater, 1987; Heaton and Kanamori, 1984). Despite the presence of abundant paleoseismic evidence for rapid coastal subsidence and tsunamis, the plate boundary remains the quietest of all subduction zones (e.g. Atwater et al., 1995; Darienzo and Peter- son, 1990; Nelson et al., 1995; Lisowski, 1991; Hyndman and Wang, 1995) and comparisons with other subduction zones (e.g. Atwater, 1987; Heaton and Kanamori, 1984). Despite the presence of abundant paleoseismic evidence for rapid coastal subsidence and tsunamis, the plate boundary remains the quietest of all subduction zones, with only one possible interplate thrust event ever recorded instrumentally (Oppenheimer et al., 1993).

Methods

Prior to the 1999 cruise, we integrated all available swath bathymetry and archive core data sets from Cascadia Basin into a GIS data base for channel pathway analysis that includes physiography, axial gradients and slope stability/slumping assessments. During our NSF sponsored cruise aboard RV Melville in July, 1999, we employed the Oregon State University (OSU) wide swath (4") coring gear to collect 44 piston cores of 6-8m length and companion trigger cores (also 4") of 2m length and seven box cores (50 x 50 x 50 cm). We collected these cores in every major canyon/channel system from the northern limit of the Cascadia subduction zone near the Nootka fault, to Cape Mendocino at its southern terminus. These included (from north to south) Barkley, Juan de Fuca, Willapa, Astoria, Rogue, Smith, Klamath, Trinidad, Eel and Mendocino canyon/channel systems (Fig. 1).

We carefully sampled channel systems using SeaBeam swath bathymetry and sidescan sonar to place the core stations. We took a full GIS workstation to sea, using the 3D ERDAS Imagine GIS system, coupled with Fledermaus 3D visualization software. Together, these systems allow us to “fly through” the site data and visualize backscatter data draped over shaded bathymetry in order to precisely locate core sites. This technique, coupled with P-code GPS and dynamic positioning, allowed us to pinpoint cores on the Cascadia margin to within 10's of meters that were surveyed only hours before.

Core sites were chosen to take advantage of known depositional segments of channels versus non-deposition or erosion by turbidity currents. Analysis of these data proved essential to successfully capturing the turbidite event record, while avoiding difficulties such as channel gravel and erosive effects that can complicate and bias the record. At sea, we analyzed all cores using the OSU Geo- tech MST system, collecting gamma (GRAPE) density, p-wave velocity and magnetic susceptibility series for each core. These data are invaluable for correlation of individual turbidite events in our Cascadia work (Fig. 3). Correlation of major stratigraphic events is based on the first occurrence of Mazama Ash in turbidites (Nelson et al., 1968, 1988) and on the approximate onset of Holocene sediment deposition as determined by a dominance of radiolarian fauna in hemipelagic sediment (Nelson, 1968; 1976; Duncan et al., 1970).

Accelerator mass spectrometer (AMS) radiocarbon ages provided by the Lawrence Livermore Laboratory in California have been determined from planktonic forams deposited in the hemipelagic sediments that underlie turbidite beds. Because previously deposited planktonic forams can be reworked when they are entrained in the turbidite deposit, we sieve the >0.062 mm sand fraction and carefully hand-pick planktonic forams only from the 0-6 cm of hemipelagic sediment below the base of the turbidite and above any underlying turbidite tail deposit that may have resedimented microfauna. Raw AMS radiocarbon ages have been reservoir corrected and converted to calendrical years (cal yr) by the method of Stuiver and Braziunas (1993). Though there may be a time-varying surface reservoir age (Southon et al., 1990) this has not yet been demonstrated. For example, the AMS age of the uppermost turbidite in our BX1 (Mendocino Channel)(Fig. 2) gives a zero age after reservoir correction and conversion to calendar years. This sample has Pb210 activity, indicat- ing an age of less than 100 years (C. Nittrouer, University of Washington pers. comm. 1997), and suggesting that the applied reservoir correction is correct within at least 100 years.

Determination of Stratigraphic Datums

The Holocene stratigraphy of submarine channels along the Cascadia margin has long been known to exhibit numerous turbidite events including excellent turbidite marker beds that contain Mazama Ash (MA) from the eruption of Mt. Mazama forming Crater Lake, Oregon (Nelson et al., 1968). The calendrical age of the eruption of Mt. Mazama has recently been re-dated at 7627±150 cal. yr BP from the Greenland Ice Core (Zdanowicz et al.,
Figure 3. (A) Shipboard core photo showing color changes at the approximate Holocene (olive gray) to Pleistocene (gray) boundary (H/P) in the central core section. This biostratigraphic boundary with an average age of 12,300 calendar years BP in Cascadia Basin actually is defined by the dominance of radiolarian fauna in the olive gray clay and foraminiferian fauna in the gray clay, whereas the Holocene time stratigraphic boundary is arbitrarily defined as 10,555 calendar years BP for the conversion of 10,000 BP. Above the H/P boundary, turbidite event beds are marked with white buttons and below the boundary in the core section to the left, Pleistocene ice-rafted cobbles are present. (B) Core sediment logger data showing sample gamma-ray density and magnetic susceptibility logs that were used to help identify and correlate the turbidite events shown in the middle and right core halves and T 1 to T 10 in the logs.

1999). The Mt. Mazama eruption airfall was distributed northeastward from southern Oregon mainly over the Columbia drainage and some of the coastal rivers. From these rivers, MA was transported to temporary depocenters in canyon heads of the Cascadia continental margin, similar to the distribution of Mt. St. Helens ash following the 1980 eruption (Nelson et al., 1988). Apparently, earthquake triggered slides resulted in turbidity currents that funneled margin sediment with MA into Cascadia Basin canyon and channel floor depocenters.

In archive cores and new 1999 cruise cores, we have reexamined the post-Mt. Mazama turbidite stratigraphy from Cascadia Basin channels utilizing high-resolution AMS radiocarbon ages. About 100 years after the Mt. Mazama eruption, the first post-MA turbidite event occurred at 7530 cal yr BP, based on an average age from cores in Cascadia, Astoria and Rogue channels (Figs. 4, 5, 6).

Our new AMS radiocarbon dates also constrain the age of faunal and color changes that are found at the approximate Pleistocene to Holocene stratigraphic boundary (H/P) (Fig. 3). This H/P boundary can then be used to establish a datum in cores with no Mazama ash occurrence and few or no AMS ages. This change was most rapid during the period from 11-13,000 yr BP, and corresponds to a sharp decline in terrigenous sedimentation, climatic change, and shifts from grasslands to forest onshore (Gardner et al., 1997; Karlin et al., 1992). In hemipelagic sediment, this Pleistocene to Holocene change results in higher organic carbon and lower carbonate content, a clear shift from Pleistocene foramin to Holocene radiolarian faunal dominance that is not geographically dependant, and a marked color change from grey to olive green that is pervasive in the Cascadia Basin cores (Fig. 3) (Nelson, 1968 and 1976; Duncan et al., 1970; Karlin et al., 1992). New ages in cores from the Barkley Cascadia, Astoria, Base of Slope, and Rogue channel systems, show that the average age of the Holocene onset (H/P) is about 12,300 cal yr BP.
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Figure 4. Graphic log for piston core 24 from Cascadia Channel showing (in red) the first occurrence of a turbidite bed containing Mt. Mazama volcanic glass (25-75%)(MA). The other post-MA turbidite bed events are numbered from 1-12, and calendar year BP ages derived from AMS radiocarbon ages are shown for selected events. These ages are used to derive the recurrence intervals (Tr) for turbidite events. Ages for events 13 and 9 were obtained from archive core samples of Griggs and Kulm (1970). The core location is shown in Figure 2.

Figure 5. Graphic log for piston core 3 from Astoria Channel showing (in red) the first occurrence of a turbidite bed containing Mt. Mazama volcanic glass (25-75%)(MA). The other post-MA turbidite bed events are numbered from 11-12, and calendar year BP ages derived from AMS radiocarbon ages are shown for other selected events. Ages for events 13 and 9 were obtained from archive core samples of Nelson (1976). The core location is shown in Figure 2.
near the base of the continental slope (Figs. 1, 4, 5, 6) (Goldfinger et al., 1997).

Cascadia Turbidite Systems

Definition of Turbidite Systems and Pathways

A wide variety of modern turbidite systems are found in present-day oceans and lakes, and even within one basin system such as Cascadia Basin. No single "fan model" describes this variety of turbidite systems with their differing sand-prone reservoir architectures. However, basic end-member types of modern siliciclastic turbidite systems have been outlined, and a nomenclature assigned, which we will use throughout this paper (Table 1) (Nelson et al., 1991). Base-of-slope sand-rich aprons are defined as those small-scale, (< 10 km) wedge-shaped turbidite systems abutting the base of slope that do not have significant channel development that is detectable in seismic profiles, side-scan mosaics, or swath-bathymetric maps (Table 1). Submarine fans are defined as those turbidite systems with significant channel development that funnels sediment into depositional lobes. Both aprons and fans can vary from mainly mud-rich to sand-rich systems, but in the tectonically active margin setting of Cascadia Basin, the dominantly mud-rich systems are not as common as they are in passive margin settings. Another turbidite system commonly found in active margin settings is the extensive, tectonically-controlled, deep-sea or mid-ocean channel system. Deep-sea channel systems are fed by multiple tributary canyons, and extend for hundreds to thousands of kilometers across a basin floor and eventually connect with abyssal-plain fans. A wide variety of feeding canyons, connecting axial channels and channel-levee complexes also are parts of turbidite systems or may make up a partially-developed turbidite system itself. Large-scale sediment wave or dune fields associated with canyon mouths and proximal channels have also been observed in the recently developed swath-bathymetric and sidescan mosaics of basin floors. All of these basic patterns of turbidite systems are found in Cascadia Basin (Table 1).

In a tectonically active setting like Cascadia Basin, folding, faulting and extensive sediment failures can disrupt canyon and channel pathways of turbidite systems. Utilizing our GIS data base for Cascadia Basin, we can determine which are the open and active pathways, where systems disrupted and what are the resulting changes to the sand-prone facies. The MA and H/P marker beds are
### Variations of Modern Turbidite Systems Along the Subduction Zone Margin, Cascadia Basin: Implications for Turbidite Reservoir Beds

<table>
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<tr>
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#### NON-CHANNELLED DEPOSITIONAL BODIES

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#### CHANNELS

**Channel Apron**

- **Fan**: Small point source of a local river or littoral drift cell.
- **Mud-Rich Fan**: Large point source of a single major mud-rich river or deep-sea channel.
- **Sand-Rich Apron**: Line source of multiple debris bodies, typically in terminally active basins.
- **Mud-Rich Apron**: Line source of multiple failures, typically on colluvial slides of muddy delta fronts and nature passive margin.

#### CONNECTING CHANNELS

- **Lateral delta**: Linear, depositional channel without significant local fan and delta.
- **Channel floor**: Coarse-grained sedimentary deposits. Levees: sand-rich thin-beded overbank turbidites.

#### DEPOSITS

- **Slipper Canyon**: Slope mass-wasting or river.
- **Deep-sea Channel**: Multiple canyons and connecting channels. Temporally-controlled basin channels that connect downstream to fans or abyssal plains.

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**Table 1. General types of modern siliciclastic turbidite systems.**

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a key for determining a complete sequence of Holocene turbidite beds and an active turbidite channel pathway up to the present time. Irregular axial gradients of channels caused by fault offsets and mass transport deposits together with missing MA turbidite sequences are keys to a disrupted channel system (Figs. 7, 8). As individual Cascadia turbidite systems are described these pathway features will be pointed out.

Cascadia Deep-Sea Channel

The Washington continental margin is characterized by a number of submarine canyons (Juan de Fuca, Quinault, Grays, and Willapa) that follow irregular pathways through the accretionary folds (Fig. 7) of the deformation front along the continental slope. (Figs. 1, 2). These tributary canyons join at the southern end of Nitinat Fan on the continental rise and form Cascadia Channel, a deep-sea channel that crosses Cascadia Basin, then incises through the Blanco Fracture Zone and continues hundreds of kilometers into Tufts Abyssal Plain. This turbidity-current pathway traversing 1000 km of Cascadia Basin has remained open throughout the late Quaternary to the present as shown by the occurrence of the 13 post-MA turbidites throughout the pathway in all recent cores we have collected (Fig. 4) and those previously examined by Griggs (1968).

Although a continuous sequence of Holocene turbidites has deposited along the Cascadia Channel floor (Fig. 4), it remains as a mainly erosive and high relief channel with poorly developed levees throughout Cascadia Basin (Griggs and Kulm, 1970; Nelson, 1976). The main deposits of Cascadia Channel probably are found on Tufts Abyssal Plain, which has not been studied. Recent research suggests that large volumes of sand from the late Pleistocene Missoula Floods of the Columbia River may have been funneled through Cascadia Channel pathways beyond Tufts Abyssal Plain to deposit 100 m of massive sands in Escanaba Trough at the far southwest corner of Cascadia Basin (Brunner et al., 1999). In its main pathway in Cascadia Basin, however, Cascadia Channel depositional architecture consists of mainly tributary channel-levee complexes that are most depositional in the northern Nitinat Fan region and mainly erosional bypass channels in central Cascadia Basin. In sum, Cascadia Channel is a deep-sea channel type turbidite system that has no submarine fan architecture with depositional lobes and consists of mainly channelized facies in Cascadia Basin (Table 1) (Griggs and Kulm, 1970).

Figure 7. Three dimensional swath bathymetric image looking southward from Astoria Canyon mouth over the Oregon continental slope and Astoria Fan. The “hot” red and yellow colors highlight the shallower depths of the accretionary folds in the lower slope. The “cool” blue green and blue colors highlight the channel pathways of Astoria Fan. Red lines trace the axial pathway of the Astoria and Base of Slope channels whose axial gradients are shown in Figure 8. The numbers in red circles represent the number of post-Mazama turbidites at each core location in Astoria Channel. Cores with 7, 6, 5, and 3 events are each located in progressively downstream channel splays and show the downstream loss of post-Mazama turbidites. The distance from the Astoria Canyon mouth to the Astoria Channel core site with 3 Post-Mazama turbidites is about 150 kilometers.
turbidites per unit time (Fig. 4), during the late Pleistocene, coarser-grained (Griggs, 1968) cyclic generation of turbidity currents (Adams, 1990). In our Holocene history dominated by channel deposition with apparent continuous gradients into several channel splays (Fig. 7) and down the main channel (Fig. 8). This pathway down the Base of Slope Channel is blocked by a lower slope slump and then a fault offset. This can be shown in the axial gradient of the Slope Base Channel that is disrupted at each feature and by the fact that the post-MA turbidites are present upstream of these channel gradient disruptions, but not downstream.

Astoria Channel and its splays show the classic u-shaped fan valley floor and levee morphology (Fig. 7) and “gull-wing” seismic stratigraphy of fan channel-levee complexes (Nelson et al., 1987). Near the canyon mouth, Astoria Channel has maximum relief of about 200 m and it gradually becomes shallower with relief of about 10-20 m by the lower fan, as channel width decreases from 5 km to 0.5 km from upper to lower fan channels (Nelson, 1976). The lower fan-shaped morphology extends to approximately 44°N (Figs. 1, 2). From here southward until the Gorda Ridge and California border at 42°N, the channels gradually merge into an extensive basin plain with flat-lying reflectors (Nelson et al., 1987). An important morphological factor to note for both the extensive Cascadia and Astoria channel systems is that there is a dominant north to south orientation of all channels parallel to the Cascadia Basin slope rather than perpendicular to the margin as most submarine fan models indicate. Both systems begin at a slightly oblique northeast to southwest orientation to the margin and then extend directly north to south through their central portions. The channels then turn to the southwest in their distal portions because of tectonic control in the case of Cascadia Channel and probably because of diversion around the local turbidite systems in south Cascadia Basin in the case of Astoria Channel (Fig. 1). These margin parallel channel systems probably result in part from the tectonic control of the subduction zone, particularly in the case of the tributary canyons for Cascadia Channel where pathways are diverted by accretionary prism fold belts along the deformation front (Figs. 1, 7). In the case of the Astoria Fan channels, there is a leftward migration of channels, becoming more parallel to the slope through time and ending in the most recently active Astoria and Base of Slope channels. This migration is caused by corollis forcing of turbidity currents and the development of higher relief.
right hand levees that we observe in fan channels in the northern hemisphere (Fig. 7) (Nelson et al., 1970; 1987).

Although the surface morphology of Astoria Channel ends abruptly at 44° N, in seismic profiles, we can trace a major subsurface continuation of this channel near the base of the slope until about 42° N (Figs. 1, 2) (Wolf and Hamer, 1999). There Astoria Channel turns to the south-west across south Cascadia Basin and again at 40° 30' emerges on the surface near Escanaba Trough as a channel 1000 m wide with 100 m relief. The southern portion of Astoria Channel has infilled because mega-landslides from the central Oregon continental slope have blocked and/or disrupted the channel gradients. First, the Heceta slide displaced a 100 km block of the margin onto the base-of-slope region (Fig. 1) (Goldfinger et al., 1997). About 11,000 years ago based on sedimentation rates, a second 25 km block (Fig. 1 see 44 N Slump) from the lower continental slope slid into Astoria Channel at 44° N on the northern end of the Heceta megaslide. This history of blockage of Astoria Channel is verified by a seismic profile showing a 10 m hemipelagic drape covering the emergent channel floor near Escanaba Trough (Fig. 1) (Wolf and Hamer, 1999). We obtained a core of 5.5m at the profile location and our 1999 core penetrates halfway into this 10 m thickness. The core contains only hemipelagic sediment compared to the thick sandy turbidites that typically are found in early Holocene and late Pleistocene deposits of lower Astoria Channel upstream of the slide blockage (Fig. 5). At 5.5m we obtained an AMS age of 13, 969 cal yr BP, which prorated to the 10m total hemipelagic thickness, suggests that the Heceta Slide and initial burial age of Astoria Channel is about 25,000 cal yr BP. Prior to these gigantic landslides, the entire Astoria Fan system had the unusual morphologic feature of a major through-going channel the entire length of the fan and across the south Cascadia Basin (Figs. 1, 2).

Like the thick turbidite beds in Pleistocene deposits of lower Astoria Channel mentioned above (Fig. 5 see T21), similar marked stratigraphic changes between Holocene and Pleistocene turbidite deposits are found throughout Astoria Fan (Fig. 9). All the Pleistocene deposits of both channel and interchannel environments throughout the lower fan contain thicker and coarser-grained turbidite sand beds compared to the general hemipelagic mud drape with thin coarse silt turbidites of Holocene time (Nelson, 1976). The sand-prone Holocene turbidite deposits are restricted to only channel floor locations and these turbidites are thinner and finer grained than those of the underlying late Pleistocene deposits (Fig. 9) (Nelson, 1976). Thus, high sand to shale ratios are found in channel floor deposits of either highstand (Holocene) or lowstand (Pleistocene) sea level times whereas high ratios are found in lower fan lobe deposits only during Pleistocene lowstand times.

Figure 9. Sand:shale ratios for the late Pleistocene (left diagram) and Holocene (right diagram) deposits of Astoria Fan. Ratios are based on total thickness of gravel, sand and coarse silt turbidite beds compared to total thickness of mud beds found in each piston core taken from Astoria Fan. The sand : shale ratio of the Pleistocene deposit of each core has been calculated separately from the Holocene deposit of each core. To make a comparison with sand: shale ratios of consolidated rocks in outcrops and drill cores, the unconsolidated mud of Astoria Fan mud beds has been compacted to 1/3 of original thickness for the sand : shale ratio calculations (Modified from Nelson, 1976).
The first MA turbidite marker bed occurs throughout Astoria Channel floor and in overbank deposits of the proximal fan (Fig. 9) (Nelson et al., 1968, 1988). With the presence of this marker bed, it is possible to show changes down-fan in stratigraphy, turbidite lithology, predictable reservoir bed characteristics and turbidity-current processes. Utilizing this marker bed it is possible to show a progressive loss of the 13 post-MA turbidites down Astoria Channel as it bifurcates into several channel splays (Fig. 7). Thus, not only are the Holocene fan turbidites less robust than those of the Pleistocene, but their dispersal down fan is restricted compared to Pleistocene turbidites that were widely dispersed in the lower fan (Fig. 9). Also, the restricted downfan distribution of Holocene turbidites in Astoria Channel contrasts with that of Cascadia Channel where there is a continuous distribution of all 13 post-MA turbidites for hundreds of kilometers (Figs. 4, 5)(Adams, 1990).

The MA marker bed also can be utilized to trace lithologic changes within the dispersal of a single turbidite event throughout Astoria Fan. In the proximal Astoria Channel, the first MA turbidite bed is a coarse sandy chaotic deposit with large wood chips, shell fragments, and pebbles up to 2-3 cm diameter that contains about 20% matrix of silt and clay (Nelson, 1976). At 150 km downfan in the distal channel, the turbidity current has evolved to deposit a well-sorted, graded fine sand deposit with a few percent matrix and a complete Bouma sequence of sedimentary structures (Carlson and Nelson, 1969; Nelson, 1976). The first MA turbidite remains as a thick turbidite (20 cm) throughout the 150 km of exposed Astoria Channel, however, the correlative levee beds are thin-bedded turbidites of a few centimeters thickness, silt mean grain size, and a high content of MA volcanic glass (Nelson et al., 1968; 1988). This shows that the lateral overbank suspension flows of turbidity currents are winnowing out the finer grained material to deposit on levees, even where channel relief is maximum. These overbank suspension flows in the Holocene, however only continued into the mid-fan region even though thick channelized turbidites continued much further down both Astoria and Cascadia Channels (Figs. 4, 5, 9).

Use of the MA marker bed and our detailed sampling in 1999 also provided new information about bed continuity and potential reservoir characteristics laterally across and distally down channel floors. As, emphasized previously, the floor of Cascadia Channel has complete continuity of all 13 post-MA turbidite beds for hundreds of kilometers down the tributary canyons and channel (Adams, 1990). In contrast, Astoria Channel has variable turbidite bed continuity, particularly in the channel floors of canyon mouth and inner fan. In some proximal locations, the entire 13 post-MA deposits are missing and in other sites, parts of the sequence are present. The best preservation of the entire 13 post-MA beds is found in lower canyon walls (Carlson, 1968) or channel floor terraces that are slightly elevated above the deepest thalweg. These data show that cut and fill processes are very active in the canyon mouth region and that lateral bedding continuity is poor. Preservation of bed continuity is better slightly down stream from the canyon mouth area, but then as described previously, many of the 13 events begin dropping out until only 3 of the 13 post-MA beds are present on the floor of Astoria Channel in the lower fan (Fig. 5). Thus, in outer fan channels, there is good bed continuity in the thick older beds, but fewer beds are deposited per unit time during the Holocene.

**Rogue Base of Slope Apron**

The continental slope off the mouth of the Rogue River in Southern Oregon has a well-developed submarine canyon with multiple canyon heads (Figs. 1, 2, 10). On the Cascadia Basin floor in front of the canyon mouth, no significant channel is evident at the surface in swath bathymetric data (Fig. 10) or in 3.5 kHz high resolution seismic profiles (Wolf and Hamer, 1999). A morphologic apron of about 5-10 m thickness and 2 km diameter, however, is found in front of the canyon mouth. The apron has seismic reflectors wedging and onlapping toward the base of the continental slope. This geometry, lack of significant channel development, and presence of turbidites containing Klamath terrane heavy minerals (Nelson et al, 1996) all suggest that a small sand-rich, base-of-slope apron has developed off the mouth of the Rogue Canyon (Table 1; Nelson et al., 1991).

The echo character of 3.5 kHz profiles, the core stratigraphy and the turbidite bed lithology, also all suggest that a sand-rich lithology with good lateral bed continuity is developed throughout the Rogue apron. The echo character in the first 20 m shows good reflector continuity across the apron and an increasing number of reflections and stronger reflections with depth (Wolf and Hamer, 1999). The latter characteristics indicate increasing thickness and coarseness of turbidite beds at depth in the apron (Damuth, 1980). Similarly, the oldest late Pleistocene turbidite beds (T22 to T25) become thicker (e.g. T25 = 35 cm) and show increasing sand shale ratios with depth compared to the overlying Holocene turbidite bed sequence (Fig. 6). For example, the Pleistocene ratios are 3.7:1, whereas Holocene ratios are 1:3. Our 2 new piston cores and 2 previous archive cores (Duncan, 1968) taken throughout the Rogue apron contain the complete stratigraphic sequence of 13 post-MA Holocene turbidites and the EH boundary with a total of 21 turbidites deposited during the Holocene. Similar stratigraphic sequences and turbidite lithology in
Figure 10. Three dimensional land map view and swath bathymetric image showing the Rogue base of slope apron, canyons and river drainage basin. The “hot” red and yellow colors highlight the on-land locations of the Cascade Mountains with Mt Mazama on the eastern margin of the map, and the Rogue River drainage basin extending from the Cascades to the coast. The “cool” green and blue colors highlight the offshore continental shelf and slope, and the Cascadia Basin floor respectively. Note that no channels are evident on the Base-of-Slope Apron. The general location of the Rogue canyons and apron are shown on Figures 1 and 2.
all Rogue apron cores suggests that all 21 Holocene turbidite beds are correlative and that lateral bedding continuity is excellent within this small sand-rich, base of slope apron.

**Trinidad Canyon, Plunge Pool, and Sediment Wave Turbidite System**

Trinidad Canyon feeds into an unusual turbidite system and itself has an unusual morphology and setting for a Cascadia Basin system (Figs. 1, 2). The canyon is associated directly with the small local Trinidad River, but derives most of its sediment from the Eel River to the south during sea level highstands (Mullenbach and Nittrouer, in press). The Eel River has a small drainage, but high sediment load nearly the same as that of the Columbia River. Recent detailed studies show that 70% of the Holocene mud from the Eel is transported obliquely northwest across the shelf and down Trinidad and Eel Canyons. The combined effect of the mud deposited in Trinidad Canyon and structural setting has resulted in slope instability and erosion of a wide (50 km) amphitheater-shaped middle to upper canyon with extensively gullied multiple canyon heads (Wolfe and Hamer, 1999). This contrasts with the other Cascadia margin canyons that are narrow (10 - 20 km) and incised across the slope (Fig. 1). The mid-upper Trinidad Canyon, however, funnels downstream into a narrow canyon of a few kilometers width that crosses an unusually steep lower continental slope (12.5°) (Fig. 11).

At the base of the continental slope the canyon feeds into a plunge pool that is 5.5 km in diameter and 50-80 m deep (Fig. 11B). Radiating out from the plunge pool, a sediment wave field extends for about 17 km and exhibits wave lengths of 3-5 km and wave heights of 20-50 m. No channel levee complexes are formed or connected to the canyon mouth (Fig. 11A). Studies of some other smaller canyons on the Cascadia and New Jersey continental margins suggest that when there is a large break in slope that exceeds 5°, then plunge pools form at the base of the continental slope in the canyon mouth region (Lee and Taling, 1998). Except for these observations, little research has been completed on the hydrodynamic processes involved in the formation of plunge pool and sediment wave deposits, although several modeling and wave tank experiments are in progress (Parker, G., Pratson, L., and Syvitski, J. oral communications, 2000).

This plunge pool and sediment wave field complex is a turbidite system since graded sand turbidite beds are found both in the plunge pool and in the sediment waves (Fig. 12A). The stratigraphy shows 25 turbidite beds deposited on the pool floor since the H/P boundary of about 12,300 y BP. We have no MA marker bed in Cascadia Basin south of the Rogue River, but the H/P datum shows that during the Holocene, the average turbidite event periodicity was 1 per 492 years. In the plunge pool, the average spacing is 23 to 28 cm between turbidite beds, whereas in the sediment wave field, turbidites occur at 55 cm and 110 cm with an even spacing of 55 cm. Although the gradational very fine sand to silt turbidite bed thicknesses of 5-10 cm are the same in both environments, turbidite events only occur half as often in the sediment waves compared to the plunge pool. The sand:shale ratios for the pool floor are 1: 3 from 0-2 m of the late Holocene deposits and 1:2 for 5 to 7 m in the early Holocene deposits. Ratios in the upper 1.3 m of a sediment wave are 1:8, again reflecting the low rate of turbidite deposition. For these calculations from modern unconsolidated sediment, we assume that the mud intervals are compacted to one third of original thickness as has been done for the Astoria Fan sediment (Fig. 9) (Nelson, 1976).

Although fewer turbidites are found in the sediment waves, the character of the turbidites is similar in all environments. Individual turbidites from both pool and wave deposits exhibit unusual double or triple sand pulses within each bed, compared to the single basal sand found in turbidites of the other Cascadia Basin turbidite systems. The 5-10 cm thick turbidites in the Trinidad system typically have two or rarely 3 distinct 1-3 cm thick, cleaner sand pulses that are several cm apart. We suspect that the tributary canyon heads that are spaced up to 50 km apart (Figs. 1, 2) (Wolf and Hamer, 1999) and apparent earthquake triggering (Nelson et al., 1996; 1999; Goldfinger et al., 1999) contribute to these unusual turbidites. We speculate that travel distances to a pool or wave depositional site from the different tributary canyon heads as well as triggering times vary as the earthquake waves travel along the subduction zone. Thus, several turbidite pulses will arrive at each depositional site caused by the combination of different travel distances and sequential earthquake triggering from each canyon source.

The apron-like geometry of the Trinidad turbidite system is somewhat like the Rogue system (Figs. 10, 11A). However, the large plunge pool in the Trinidad canyon mouth and lack of lateral bed continuity between the pool and waves set it apart as a different type of turbidite system than the Rogue apron. Within the Trinidad plunge pool, 3 cores all show 25 post H/P events (Fig. 12A). In the sediment wave fields, limited data shows that only about half of the events continue from the plunge pool across the sediment wave area. Thus, there is discontinuous lateral bedding in the Trinidad pool and wave system compared the Rogue apron where there appears to be lateral bedding continuity from the apron apex to the distal apron.
Figure 11. (A) Three dimensional swath bathymetric image looking down on the Trinidad Canyon, plunge pool and radiating sediment wave field in the base-of-slope environment in southern Cascadia Basin. (B) Axial profile downstream across these proximal to distal environments.
Figure 12. Graphic log for piston cores 36 and 41 from Trinidad plunge pool and Eel Channel/lobe showing numbered turbidite bed events found above the Holocene/Pleistocene boundary (H/P). Average age for this boundary is derived from AMS calendar year BP ages obtained from cores taken in Juan de Fuca Channel, Cascadia Channel, Rogue apron and Base of Slip Channel (Goldfinger et al., 1997) turbidite systems (Figs. 1, 2, 4, 6). The H/P boundary age is used to derive the recurrence intervals for turbidite events. The general core locations are shown in Figure 2, and the specific locations are shown in the detailed bathymetric images of Figures 11 and 13.
The geomorphology and lateral bedding continuity of the Rogue and Trinidad turbidite systems also offers some clues about the history of sedimentary processes in these systems. On the Rogue apron, the Holocene turbidite currents appear to be depositing continuously across the apron with little cut and fill deposition. During the Pleistocene, stronger turbidity currents appear to have deposited thicker turbidites on the apron (Fig. 6). During the Holocene, the Trinidad pool at the canyon mouth appears to trap most of the turbidite deposition and permit only about half of the turbidite events to deposit in the sediment wave field. The stronger turbidity currents of the Pleistocene appear to have eroded the Trinidad pool and had greater deposition in the sediment wave field compared to the Holocene currents. Confirmation of this idea will require longer cores that penetrate into Pleistocene deposits (Fig. 12A).

**Eel Canyon, Plunge Pool, Sediment Wave, Channel and Lobe Turbidite System**

Morphology of the Eel turbidite system in the proximal region is somewhat similar to the Trinidad system (Figs. 1, 2, 11, 13). At the base of the continental slope, the steep (10°) narrow (<10 km) canyon feeds into an irregular plunge pool that is about 5 km in diameter and 50 m deep (Fig. 13). Radiating out from the plunge pool, a sediment wave field extends for about 25 km and exhibits wave lengths of 2.5 to 5.5 km and wave heights of 30-80 m. This sediment wave field has larger waves and is longer than that in the Trinidad pool and wave system. Also, an irregular channel threads its way through the sediment waves and is connected to a well-developed channel-levee complex downstream that has a channel floor width of 1.8 km and relief of 25 m on the higher right-hand levee (Wolf and Hamer, 1999). This channel-levee complex extends for 25 km beyond the sediment wave field and then forms a ponded lobe in a local graben structure.

The sand-rich Eel turbidite system has numerous, but thin graded sand turbidite beds in all canyon mouth, plunge pool, sediment wave, channel and lobe environments (Fig. 12B). We have observed 14 turbidite beds on the pool floor (core not shown) and 51 in the lobe that have been deposited since the Holocene to Pleistocene (Fig. 12B). We have observed 14 turbidite beds on the pool floor (core not shown) and 51 in the lobe that have been deposited since the Holocene to Pleistocene (Fig. 12B). The H/P datum shows that during the Holocene, the average turbidite event periodicity in the lobe, that has the most complete event record, was 1 per 246 years. In the Holocene deposits of the plunge pool and sediment wave field, the average spacing is 18 cm between turbidite beds whereas in the lobe it is 9 cm or essentially twice as often as in the plunge pool. The floor of the Eel Canyon mouth has only Pleistocene deposits and the spacing there (20 cm) is similar to the other proximal turbidite environments.

The sand-shale ratios in all the Holocene and Pleistocene deposits for the Eel system are relatively high compared to the Trinidad system; the canyon mouth is 1:1 and the pool floor is 1:2:1 in the early Holocene, and 1:2:4 in the Pleistocene; the channel levee 24 km from the canyon mouth is 1:3.2, and the lobe 50 km from the mouth is 1:1.2 from 0-1.5 m of the late Holocene deposits and 1:8:1 from 300-350 cm in the early Holocene deposits (Fig. 12B). Similar to other turbidite systems in Cascadia Basin, the stratigraphy shows a greater proportion of turbidite beds in the latest Pleistocene and early Holocene deposits (i.e. ratios of 1:1 or greater) relative to late Holocene deposits (i.e. ratios of 1:1.5 or less). The Eel plunge pool environment is an exception to this stratigraphic trend because the Pleistocene deposits are represented by thin-beded turbidites and the early Holocene deposits contain thick sand beds as shown by the sand-shale ratios above.

We have not been able to sample what appear to be thick sand turbidites on the floor of the central Eel channel-levee complex. The 3.5kHz seismic profiles show high reflectance and no penetration (Wolf and Hamer, 1999) that is typical of thick channel beds (Damuth, 1980). Neither gravity nor box cores have been able to penetrate what appear to be thick channel floor sands, again suggesting the sand-rich nature of the Eel system. The levee deposits of the Eel Channel contain thin-beded turbidites that are typical of such environments (Nelson and Nilsen, 1984).

The thickest Eel turbidite beds (10-15 cm) are found in the canyon mouth floor and early Holocene pool deposits, although we have not sampled what may be thickest beds in the downstream channel floor. The lobe turbidites are thinner than those in more proximal environments, but are consistently 3.5 cm thick throughout the Holocene deposits (Fig. 12B). Also, as the sand-shale ratios indicate, turbidite thickness varies with time, because thin beded turbidites are generally characteristic of the late Holocene deposits, although there is the change to thin-beded turbidites in Pleistocene deposits of the plunge pool. The change from thin Pleistocene to thick early Holocene turbidites in the pool may indicate a change from stronger bypassing and eroding turbidity currents during the Pleistocene compared to more depositional currents in the pool during the Holocene. These data from the Eel pool appear to support the same hypothesis we have proposed for the Trinidad pool and wave system.

The lateral continuity of turbidite beds varies with the different geomorphic environments and with time in the Eel turbidite system. Similar to the Astoria Canyon mouth and proximal channel environments, the proximal canyon mouth and plunge pool floor of the Eel system show poor
Figure 13. (A) Three-dimensional swath bathymetric image looking down on the Eel Canyon, plunge pool, radiating sediment wave field, downstream channel-levee complex and lobe in the base-of-slope environment in southern Cascadia Basin. (B) Axial profile downstream across these proximal to distal environments.

Turbidite bed continuity that suggest cut and fill environments. Holocene deposits have bypassed or been eroded out of the canyon floor. Thin-bedded Pleistocene turbidites of the pool floor suggest sediment bypassing compared to early Holocene deposition of thick turbidites. Significant turbidite bypassing of the pool environment also is shown by the deposition of only 14 turbidite events during the Holocene compared to 51 events deposited in the lobe during the same time (Fig. 12B). Similarly, the 18-20 cm wide spacing of turbidite events in the canyon floor, pool, and levee environments compared to the 9 cm spacing of events in the lobe indicates loss of events and less turbidite bed continuity in all of these environments compared to the lobe.
Figure 14. (A) Graphic log for box core BX 1 from Mendocino Channel showing numbered turbidite bed events and calendar year BP ages derived from AMS radiocarbon ages. These calendar ages are used to derive the recurrence intervals for turbidite events. (B) Gloria sidescan image from central Mendocino Channel with interpretive drawing showing the detailed location of the BX1 sample in the channel. Archive core samples for the ages as well as the Gloria sidescan image and interpretation were obtained from Cacchione et al. (1996). The general core location is shown in Figure 2.

Mendocino Channel

The Mendocino Canyon feeds the Mendocino Channel that lies at the base of the Mendocino Escarpment (Figs. 1, 2). The escarpment topography and the meandering channel pathway have been controlled by the extensive tectonic activity of the Mendocino Triple Junction (Cacchione et al., 1996; Wolf and Hamer, 1999). Even with the extensive tectonic disruption, this turbidite pathway has been more active than any other Cascadia Basin turbidite system and at its proximal end has canyon heads that erode to within 1 km of the shoreline. After meandering along the escarpment, the distal end of the channel turns northwest into the Gorda Basin where it may coalesce with the Eel turbidite system. The Mendocino Channel does not exhibit proximal plunge pools or distal lobe development in any of the
3.5 kHz profiles downstream (Wolf and Hamer, 1999), and thus it is a connecting channel levee complex type of turbidite system (Table 1). However, in the previous interpretation of Cacchione et al., 1996 that is shown in brown on 14B, they have drawn a lobe feature at a channel termination, whereas this is a chaotic failure surface on the northern levee of the through-going channel levee complex (Wolf and Hamer, 1999). We have only sampled this narrow channel (0.5 km width) in its central area of tight meanders approximately 20 km downstream from the canyon mouth.

The sand-rich Mendocino turbidite system has the highest quantity and most rapid rate of deposition of graded sand turbidite beds in any Holocene deposits of Cascadia Basin. Because no MA turbidite bed or H/P datum was sampled, only Holocene deposits of younger and older ages can be compared. The stratigraphy reveals 4 turbidite beds deposited during the past 260 cal. yr B.P. (14A). These AMS ages show that during the latest Holocene, the average turbidite event periodicity in the central channel was 1 per 65 years, more rapid than any other location in Cascadia Basin (Fig. 15B). A preliminary age, at a greater depth in a 1999 piston core (not shown) at the same location as BX 1, suggests that the frequency of events may be even greater at 1 per 40 years.

In the latest Holocene deposits of central Mendocino Channel, the average spacing is 9 cm between turbidite beds, whereas at 3.5 to 5 m depth, it is 19 cm. The average thickness however is 3.8 cm in the early Holocene sediment and 2.8 cm in the late Holocene sediment. Similarly, the sand-shale ratios in the older deposits are relatively higher (2.5:1) compared to the younger deposits (2.1:1), although both of these Holocene sand-shale ratios are the highest observed in the Holocene deposits of any turbidite system within Cascadia Basin.

The thickest turbidite bed (12 cm) in Mendocino Channel is found in the older deposits, although turbidites of about half that thickness occur in the youngest deposits (Fig. 14A). The variation of turbidite thickness with time also is shown by the sand-shale ratios of the late Holocene deposits, which are less than those of the older Holocene deposits. Because we only have samples from the central channel and no stratigraphic datums, we do not have data about the lateral continuity of turbidite beds. The high sand-shale ratios and lack of any observed cut and fill in the turbidite deposits suggests that lateral and vertical continuity is good in the central channel region.

Seismic Triggering and Rate of Turbidite Events

Adams (1990) observed that 13 post-MA turbidite events correlate for hundreds of kilometers along Cascadia Channel from the tributary canyons at least to the Blanco Fracture Zone (Fig. 15A). Because the channel tributary canyons of Juan de Fuca, Willapa, Grays, and Quinault Canyons (Fig. 1) contain equal or a few more post-MA events, he reasoned that if these events had been independently triggered events with more than a few hours separation in time, cores taken below the confluence should contain from 26-31 turbidites, not 13 as observed. The importance of this simple observation is that it demonstrates synchronous triggering of turbidite events in tributaries, the headwaters of which are separated by 50-150 km. Adams (1990) also noted that 13 events are present at the mouths of Astoria and Rogue Canyons off northern and southern Oregon, suggesting 13 plate-wide Holocene ruptures (Fig. 15A).

Are these events triggered by earthquakes? Adams (1990) suggested four plausible mechanisms for turbidite triggering: 1) storm wave loading; 2) earthquakes; 3) tsunamis; 4) sediment loading. Sediment loading, the most common trigger on passive margins, fails to explain any sort of synchronicity of events even short distances apart, and thus it is unlikely that a self-triggering mechanism could generate similar numbers of turbidites for 7500 years along 700 km of the margin. The three remaining mechanisms could trigger regional events. Storm wave loading is an unlikely cause in Cascadia, where, although deep water storm waves are large, the canyon heads where accumulation occurs are at water depths of 150-200 m. These depths are at the maximum possible depths for even the gentlest disturbance by storms with maximum significant wave heights of 20 meters, making this an unlikely trigger. Although the Oct. 12th storm of 1962 with winds > 300 km/hr and the catastrophic flooding of 1964 and 1998 El Nino winter storms have occurred, no turbidite event has been deposited during the past 300 years along the northern 700 km of the Cascadia margin (Fig. 4). Tsunamis may also be a regional trigger, however, the 1964 Alaska Mw 9.2 event did not trigger a turbidite observed in any of the Cascadia Basin cores, although it did extensive damage along the Pacific Northwest coast. Adams arguments (1990) as well as storm and tsunami wave evidence suggest that plate-wide earthquakes, though not unique, are the most likely regional trigger in Cascadia Basin.

We have verified Adams (1990) hypothesis, from Vancouver Island to the Rogue Canyon, by carefully collecting multiple cores in all Cascadia Canyon and Channel systems in 1999 (Fig. 15B) (Goldfinger et al., 1999; Nelson et al., 1999). Previous evidence for 13 post-MA events in Astoria Channel was weak and suggested plate segmentation rather than plate-wide earthquakes (Nelson et al., 1996). In archive cores, only 3 post-MA events were found in middle and lower Astoria Channel, which appeared to
contradict Adams (1990) hypothesis for 13 events. In new cores, we find a progressive loss of turbidites from 7 to 6 to 5 events at each successive downstream channel splay in upper Astoria Fan (Figs. 7, 15A). This down-channel loss of events, resulting in only 3 events in the mid-Astoria Channel, explains the previous apparent contradiction.

In sum, for 700 km along the Cascadia subduction zone, the most consistent interpretation is that 13 post-MA events occurred (Fig. 15A). Assuming event 13 took place 7500 cal yr and event 1 took place 1700 AD (Nelson et al., 1995; Satake et al., 1996), 12 turbidite events have occurred during 7200 years or a mean recurrence time of 600 years (7500-300/2= 600) (Fig. 15B). A 600 year turbidite recurrence interval also is observed between the MA and HP datums in the Astoria, Rogue and Klamath turbidite systems (Figs. 5, 6). The same periodicity between most events in all cores also is suggested by the consistent thickness of hemipelagic sediment representing about 600 yr between turbidite beds (Fig. 4) (Adams, 1990). The similarity in number of events that has now been verified in multiple new cores in each canyon/channel system is remarkable, and precludes virtually all non-earthquake triggers. We can think of no other plate-wide events that could cause such similar records in time and space. Thus, for the northern 2/3 of the Cascadia margin, we infer that plate-wide, large-magnitude earthquakes occurring every 600 years are the best explanation for the turbidite event record.

For the northern California margin, the record is more complex. The number of evenly-spaced Holocene turbidite
Summary of Turbidite Bed Architecture and Potential Reservoir Facies in the Different Types of Cascadia Basin Turbidite Systems

The correlative MA and H/P stratigraphic markers together with synchronous seismic triggering of turbidity-current events provides a unique opportunity to summarize the architecture and continuity of sand-prone reservoir facies for a wide variety of turbidite systems in a tectonically active continental margin. Because whole-plate seismic triggering seems to be dominant in Cascadia Basin, turbidite deposits are quite frequent and widespread, even in Holocene deposits. As seismicity and turbidity-current triggering increase toward the triple junction of Cascadia Basin (1 per 600 yr) and the progressively increasing frequency of events toward the triple junction can best be explained by seismic triggering (Fig. 15).

In general, the best development of sand-prone turbidite facies is found within channel floors, distal lobes, and proximal aprons (Figs. 4, 6, 9). The best lateral continuity of Holocene turbidite beds is found in aprons, distal channels and their depositional lobes (Figs. 4, 6, 9, 15A). As shown in Astoria Fan (Fig. 9B) (Nelson, 1976), Cascadia Channel (Griggs and Kulm, 1970), and comparison of the Trinidad and Eel plunge pools and sediment waves, the poorest bed continuity and lowest sand-shale ratios (1:9) are found in levees and sediment waves. The thickest and coarsest-grained turbidite beds and correlative highest sand-shale ratios (3:1) occur in Pleistocene deposits (Figs. 6, 9) (Nelson, 1976; Griggs and Kulm; Duncan, 1968). High sand-shale ratios (2.5:1), however, are found in early Holocene sequences.

The architecture, sand-bed continuity and depositional processes vary considerably in different turbidite systems of the Cascadia Basin. In the extensive (1000 km) Cascadia deep-sea channel, lateral continuity of Holocene turbidites on channel floor extends for hundreds of kilometers down channel (Fig. 15A) (Adams, 1990). Levee deposits, however, show only thin-bedded turbidites with poor continuity and cutout of beds laterally away from the channel floor (Griggs and Kulm, 1970). In the deep-sea channel, the high relief of channel walls and lack of channel splays result in continuous funneling of turbidity currents for great distances down channel and in extensive channel-bed continuity. In contrast, the Holocene deposits of the proximal channel floors in Astoria Fan exhibit massive cut and fill of chaotic, high matrix muddy gravelly sandy turbidites; these evolve down channel to mature well-sorted, low matrix, and thick fine sand beds (Nelson, 1976). Bed continuity, however, is poor with 10 out of 13 post-MA turbidites dropped out in the distal channel of the lower fan (Figs. 7, 15A). The bifurcations of Astoria Channel into multiple channel splays down fan (Fig. 7), in contrast to lack of bifurcation or lobe development and continuous high-relief of Cascadia Channel down stream, seem to account for the difference of bed continuity in channels of these two different types of turbidite systems. Both the lack of channel bed continuity and high Pleistocene sand-shale ratios of the lobes suggest that significant sand bypassing occurs in the fan channels, especially during Pleistocene times (Figs. 9, 15A).

The lateral continuity of turbidite bed deposition in the Rogue base-of-apron is good, probably because of the lack of channel-levee processes that disrupt bed continuity. In the Trinidad turbidite system that lacks channels, high frequency of turbidite bed deposition, high sand-shale ratios, and good bed continuity are found in the plunge pool at the canyon mouth compared to the opposite of these characteristics found downstream in sediment waves. In the Eel system, the trends of turbidite deposition, sand-shale ratios and bed continuity in the plunge pool and sediment wave environments are similar to those of the Trinidad turbidite system; however, a major channel bypasses large amounts of sand to the distal lobe that has the highest sand-shale ratios and most complete record of turbidite deposition (Fig. 12B). The Mendocino Channel, a connecting channel-levee complex, has the most rapid rate of turbidite events and highest sand-shale ratios probably as a result of its tectonic setting at a triple junction (Figs. 1, 2, 14). The progressively increasing frequency of turbidite deposition and trend to higher sand-shale ratios in turbidite systems near triple junctions may be a trend to look for in subsurface reservoirs of tectonically active margin settings.

The most important take home message is that a wide variety of turbidite systems can develop along a single basin margin setting like Cascadia Basin. Each system has its own architecture of sand-prone facies and bed continuity in different environments. Both the facies and continuity can change at different times within the same
environment as tectonically and climatically driven sedimentary regimes change. New swath bathymetry and sidescan sonar images show that a large break in slope (>50o) at the canyon mouth, results in turbidite systems that develop plunge pool and sediment wave morphologies which somewhat mimic turbine facies of channel floor and levee facies. Presence of robust turbidites in Holocene deposits and thin turbidites in Pleistocene deposits of the plunge pools suggests that during the Pleistocene pools were eroded and bypassed by turbidity currents whereas during the Holocene the pools are infilling. Even these trends, however, can be modified if the pool and sediment wave facies become drained by a channel and lobe system.

References

Dumitru, I.E., 1980, Use of high frequency (15.5 - 12 kHz) echograms in the study of near-bottom sedimentation processes in the deep sea: A review: Marine Geology, v. 38, p. 51-75.
Variations of Modern Turbidite Systems Along the Subduction Zone Margin, Cascadia Basin: Implications for Turbidite Reservoir Beds


