CHAPTER 2B

Sub-Aqueous Paleoseismology

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2B.1 Introduction

2B.1.1 Scope of the Chapter

Many of the largest earthquakes are fundamentally marine events, generated by submarine subduction zone or other plate boundary earthquakes, as well as volcano-tectonic explosions. A large proportion of the world’s population lives near coastlines, thus a high proportion of hazard from active tectonics comes from submarine fault systems and volcanic and landslide generators of tsunami. During and shortly after large earthquakes, in the coastal and marine environment, a spectrum of evidence is left behind. Onshore, land levels change with elastic unflexing of the formerly coupled plates, resulting in coastal subsidence, uplift or lateral shift, and the generation of familiar onshore paleoseismic evidence such as fault scarps, colluvial wedges, damaged trees, landslides, and offset features. If the seafloor is shaken or displaced, another suite of events may result in further geologic and geodetic evidence of the event, including turbidity currents, submarine landslides, tsunami (which may be recorded both onshore and offshore), soft-sediment deformation, as well as virtually all of the evidence normally associated with onshore faults, including direct measurement of coseismic displacement.

Offshore and lacustrine records offer the potential of good preservation, good spatial coverage, and long temporal span. Marine deposits also offer opportunities for stratigraphic correlation along the source zone, something typically difficult with land paleoseismology. Stratigraphic correlation methods have potential to address source zone spatial extent, segmentation, and because of the longer time intervals available, can be used to examine recurrence models, fault interactions, clustering and other phenomenon commonly limited by short temporal records. Offshore deposits can be investigated geologically and geophysically to define their extent, stratigraphic relationships, and timing. Detailed investigations of marine deposits at the millimeter scale is now routine, and high-resolution geophysical techniques allow subsurface mapping and correlation with core samples to delineate mass transport deposits and turbidites. In some cases, direct evidence of earthquake slip is available and can be imaged using geophysical techniques. Many deposits, however, do not have a direct physical link to their causative sources and must be distinguished from other deposits through either regional correlation, dating, or sedimentological character. Submarine deposits may include a wide range of features and structures which overlap with those of onshore deposits. This chapter discusses mostly submarine deposits of transported nature and direct fault observations. Sub-aqueous soft sediment deformation is included in Chapter 7.
2B.2 Mapping and Dating Paleoseismic Landforms Offshore

2B.2.1 Submarine Mapping and Imaging Methods

Due to the inherent difficulties of working in the submarine environment, only a few examples exist of successful determinations of slip rates and paleoseismic histories through direct observation and imaging of active faults. Much more common are rough estimates of recent activity based on morphological evidence of scarp freshness, presence of associated landslides, mud volcanoes, fluid venting, tsunami deposits, and other secondary evidence. Methods of investigation in the submarine environment include some commonality with paleoseismic and tectonic investigations onshore, particularly topographic analysis and seismic reflection profiling, while other methods are relatively unique to the marine environment such as sidescan sonar imaging and acoustic geodetics.

2B.2.1.1 Seafloor Mapping Techniques

The earliest seabed mapping tool was a lead line, a simple lead weight on the end of a line, heaved by a sailor to manually sound the depth, and also collect a bottom sample with the lead which was “armed” with tallow or beeswax to bring back some of the bottom sediment. Modern mapping technologies now mostly use acoustic techniques. Because light and electromagnetic transmission in water is generally poor, modern photographic, radar, laser, and other such techniques common to terrestrial mapping are ineffective in the marine environment except for very short range applications. The exceptions include laser line-scan imaging, which can be done from surface vessels or towed vehicles at a low altitude above the seafloor in clear conditions; towed, lowered, or vehicle mounted cameras; and bathymetric LiDAR, which can map the seafloor from aircraft in very clear and shallow water conditions (Figure 2B.1). Most seafloor mapping is performed with acoustic technologies using sound waves reflected from the seafloor to establish both elevations (bathymetry) and backscatter strength of seafloor materials, which is a function of hardness, grain size, and material properties. Early mapping began with single-beam echosounders and recording fathometers, which emit an acoustic “ping” toward the seafloor that is then reflected back to the surface. The travel time of the ping is measured and converted to water depth using the sound velocity of water (≈1500 m/s). The transducer typically emits a cone-shaped beam pattern spanning 5–40° of arc, thus the beam pattern on the seafloor becomes larger with water depth, a limiting resolution factor in most sonar mapping tools.

Multibeam Bathymetric Sonars  Modern seafloor mapping is performed with two primary tools: multibeam bathymetric sonars and sidescan sonars. Multibeam sonars operate on the same principle as the earlier single-beam sonars, but with multiple elements arranged in an array to establish a fan-shaped pattern of multiple beams that maximize across-track width (Figure 2B.2). In this way, a broad swath of seafloor can be mapped in a single pass. Typical swath widths are three to seven times the water depth, with individual beam widths of 1–2°. Ship motion (roll, pitch, yaw and heave, the vertical motion from waves) is recorded during data acquisition and removed during processing. More sophisticated systems employ “beam steering” to “form” the individual beams electronically in real time as the ship moves, keeping the fan of beams pointed downward as the ship rolls. The data are also corrected for tides and water velocity (further processing details are discussed in Blondel and Murton, 1997). The result is a bathymetric surface along the ship track, and a large area is surveyed by arranging the ships course in a series of linear tracks arranged to best image the seafloor in the prevailing weather conditions at the time of survey. In addition to bathymetric data, the strength of the returning beam can be quantified, providing
a map of reflection intensity from the seafloor. When corrected for the effect of grazing angle at the seafloor, these data provide a map of relative reflectivity, which is commonly interpreted as patterns of outcropping rock, sand and mud at the seafloor. Multibeam sonars typically use 16–200 beams arrayed as a fan with a span of 90°–150°. Frequency of these systems ranges from 12 to 400 kHz, with lower frequency systems used for deeper water applications to overcome the attenuation of high-frequency signals in the water column.

**Sidescan Sonars**  
Sidescan sonar similarly maps the seafloor in a wide swath, but instead of forming discrete beams, the entire returning signal is digitized, yielding a higher across-track resolution of returning signal strength. Because there is no angular control of the signal, these data are processed as a signal strength image, rather than bathymetry. Sidescan imagery typically consists of 2048 across-track samples per ping, yielding much higher spatial resolution (Figure 2B.3). By using two frequencies, interfering beam patterns can be processed to extract bathymetric data as well; although typically these data are not as high in quality as multibeam bathymetric sonars (Johnson and Helferty, 1990; Blondel and Murton, 1997). Sidescan sonar offers detailed imagery that is generally superior for mapping of tectonic features, and when combined with multibeam bathymetry, yields a 3D surface map that images both

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**Figure 2B.1:** (A) Example of LiDAR bathymetry. Images of the southwest coast of Puerto Rico were taken using a LiDAR ADS Mk II Airborne System. The 900 Hertz (1065 nm) A Nd: Yag laser acquired 4 × 4 m spot spacing and 200% seabed coverage. Image source: NOAA biogeochemistry products: http://ccma.nos.noaa.gov/images/biogeo/LiDAR_pr.jpg. (B) Laser Line-Scan Survey (LLS) image showing sharp boundary between sand waves (top left corner) and rippled seafloor. Dark objects in the area of sand waves are pieces of drift kelp and a Salp chain shadow is seen at bottom. Altitude of LLS system is approximately 6 m off seafloor, and swath width is approximately 8 m. Image in “B” from Yoklavich et al. (2003), reprinted by permission of the Marine Technology Society.
morphology and material properties in the area of interest. Such information is difficult to duplicate on land except in desert environments. Sidescan sonars generally operate in the frequency range ~126 to 400 kHz, with the lower frequencies used for deep water applications, and the higher frequencies used for shallow water or deep towed applications due to the attenuation of higher frequencies in the water column. Both types of systems may be hull-mounted, mounted on retractable or temporary overside poles, towed vehicles, or on autonomous underwater vehicles (AUVs), remotely operated vehicles (ROVs) and manned submersibles.

Sidescan imagery and multibeam bathymetry can be combined to produce a 3D image showing topographic details, as well as backscatter strength, analogous to radar imagery draped over subaerial topography.

Seismic Reflection Profiling  Subsurface structures are imaged in the marine environment much as they are on land, using primarily seismic reflection profiling, with ground truth from drill holes and core
samples when possible. Seismic profiling again uses sound waves, but at a much lower frequencies to
penetrate the seafloor substrate. Marine seismic profiling is a sophisticated process developed largely for
the petrochemical industry which requires both deep penetration in deep water, and high resolution.
Spatial resolution is improved through the use of large arrays of tuned sound sources, and long arrays of
towed receivers which receive each sound pulse at many different receive points, allowing “stacking” of
weak signals, as well as calculation of velocities of subsurface units. Paleoseismology generally makes
use of simpler shallower penetration systems that focus on the upper subsurface section using higher
frequencies to identify disruption of young strata, fault details, fault terminations, colluvial wedges and in
general the same types of features observed in terrestrial trenches, though not quite at the same resolution
(e.g., Seitz and Kent, 2005; Ridente et al., 2008; Brothers et al., 2009). Seismic profiling can also image
the spatial continuity of turbidites and subsurface mass wasting deposits, as well as key reflectors that can
be used to establish temporal control of structural movement (Figure 2B.3). Such systems tend to use
operating frequencies from 3 to 300 kHz, commonly from 3.5 to 20 kHz, and may sweep through a range
of frequencies in their outgoing pulses known as “CHIRP” (Compressed High Intensity Radar Pulse,
Figure 2B.4). This has the effect of increasing resolution by including high frequencies, but still
maintaining enough energy in a longer outgoing pulse to achieve penetration of the water column and
subsurface (Verbeeka and McGee, 1995). CHIRP sonars may also be used to invert the reflection
amplitude and phase data to obtain material properties such as density, porosity, and sound speed (Turgut
et al., 2002; Schock, 2004).

2B.2.1.2 Sampling Methods

Coring Tools Among the most common sampling tools in the marine environment are coring tools,
shown in Figure 2B.5. Cores are commonly taken with a device lowered from a vessel on a wire or
synthetic rope to the seabed. Common tools are the gravity corer, piston corer, box corer, Kasten corer,
and also include the vibracorer and freeze corer. Gravity corers are the simplest of devices, consisting of a barrel, a cutter at the nose, and a weight at the top end with a valve to allow expulsion of water. A simple one way “catcher” at the nose cone prevents the sample from sliding back out the core tube. Small push corers are routinely deployed from submersibles and ROVs to collect short soft sediment samples. The Kasten corer is a large diameter, usually square, variant of the gravity corer used when large sample volume is desired. An improvement to this scheme is the piston corer, which uses a piston inside the core liner that is attached directly to the lowering wire, and a trigger corer that touches down first, triggering a freefall of the main corer. As the piston corer penetrates, the piston remains near the level of the seafloor, and generates a vacuum that helps pull the sample into the tube. This method helps overcome the friction of penetration and allows much longer samples that also have reduced internal compaction of the sample. Very long piston coring systems are in use that are capable of collecting 50–70 m core samples from very deep water. Figure 2B.4: (A) Northeast oriented, fault-perpendicular CHIRP seismic profile across the Fallen Leaf Segment of the West Tahoe Fault, Lake Tahoe California. Slip during the most recent event (MRE) was up to ~3 m in the upper 30 m of sediments. The increase in offset with depth is likely due to changes in sediment compaction rather than multiple events. A slide deposit (SD) mantles the event horizon and infills the accommodation created during the MRE. Piston core 3 sampled a sandy turbidite layer that is interpreted to be the distal reaches of the same slide sampled in piston core 1. Radiocarbon dating and the presence of the 7600–8000-year-old Tsoyowata Ash place the MRE at ~4000–5000 years BP and constrain the sedimentation rate at ~1 mm/yr over the last ~8 ka. Down-section, faintly imaged strata along the hanging wall abruptly change from horizontal to a ~4° dip toward the fault. The change in dip is interpreted to have formed during older events along the WTF. (B) Enlarged section showing deformation associated with the MRE and the projected location of PC3 onto the profile (the two are offset by <30 m). Figure and interpretation. Image from Brothers et al. (2009), used with permission from the seismological society of America.
soft materials without requiring a drilling vessel. These systems use lightweight synthetic fiber ropes to overcome the limitations of heavy wire in deep water applications.

The box corer is a broad box device that captures the uppermost ~0.5–1 m of sediment, and the multicorer collects and undisturbed set of six to eight samples of the seafloor surface, including the overlying water. Two important variants include vibracoring and the freeze core. A vibracorer consists of a core barrel with a motor mounted at the head to generate vibrations in the core barrel. The vibration causes sediments in contact with the outer barrel to liquefy, making possible the penetration of sandy lithologies that would be problematic for other techniques. The freeze corer consists of a simple tube and weight, or multiple tubes filled with dry ice, liquid nitrogen, or other cold source. The tube is lowered into the sediment, which freezes to the outside of the metal tube (Hill, 1999; Kondolf and Piégay, 2003). This device can collect delicate samples from very soft sediments difficult to recover with other methods.
Drilling  Drilling from platform rigs and surface ships has for decades provided age control as well as lithologic and geophysical ground truth as a compliment to seismic reflection profiling. Commercial and academic drilling vessels can extend the sampling reach far deeper into the substrate than passive sampling techniques and can employ downhole geophysical techniques to collect geophysical and physical property data from the walls of the boreholes. Both of the current integrated ocean drilling (IODP) platforms, the JOIDES Resolution and the Chikyu have as one of their missions, investigations of earthquakes, and thus paleoseismology is a priority. The JOIDES Resolution is a conventional nonriser drilling vessel, meaning that it cannot compensate for formations under pressure at depth. The Chikyu, the first academic riser vessel, uses heavy drilling “mud” weight to balance formation pressure, allowing deeper drilling into less stable formations. Maximum drilling depth depends on the water depth, but can reach up to 8 km, though 1–3 km is more typical for the D/V Resolution, and up to 10 km for the D/V Chikyu, which was designed in part to reach the seismogenic plate boundary in subduction zones. These drilling vessels typically recover core, which is logged analyzed and sampled much as other types of cores. In addition, they have the capability to install instruments in the boreholes and to plug the borehole with a pressure instrument to record time series of formation pressure to monitor and thus reveal patterns of stress change related to earthquakes and other phenomenon (Figure 2B.6).

Other drilling devices such as the portable remotely operated drill (PROD) and the benthic marine sampler (BMS) are devices that fill the gap between coring and ship-based drilling, and are lowered to the seafloor on cables, from where they can drill and core up to ~100 m depth from ships of opportunity.

2B.2.2 Dating Submarine Structures, Landforms, and Deposits Using Paleoseismic Stratigraphy

2B.2.2.1 Radiocarbon Dating

To date submarine events, the most common technique is to date calcareous microfossils, commonly planktonic foraminifera using $^{14}$C. This technique can be used to date turbidites, submarine landslides, and other marine disturbance events by sampling the youngest material below, or the oldest material above the event. In dating turbidites, samples are commonly taken below each turbidite because the boundary between the top of the turbidite tail and the hemipelagic sediment is difficult to identify reliably and bioturbation is concentrated at this boundary (Goldfinger et al., 2008, 2009, Figure 2B.7). Sediment samples are taken to avoid visible or undetected deformation and friction drag along the core walls. Further processing details are given in Goldfinger et al. (2009).

Foraminiferal samples are dated using accelerator mass spectrometry (AMS) methods which can make use of as little as a ~1 mg carbon sample. Sensitivity tests for species-specific biases and other techniques are presented in Goldfinger et al. (2007a).

All radiocarbon ages must be calibrated to account for variability of carbon isotopes in the atmosphere. This variability has been captured as discussed in Chapter 2; however, marine ages have several additional complications, most importantly the reservoir correction. This value, representing the age of the seawater populated by microfossils used to date marine events, is a published spatially varying value specific to the locality of interest (e.g., Reimer et al., 2004). The published value is commonly derived from paired shell/wood dates that establish the age of the water in which the marine animal lived relative to stratigraphically correlated terrestrial material. The published values are almost exclusively from the twentieth century; although it is known that these values change through time (i.e., Kovanen and Esterbrook, 2002). Time variation of the reservoir age is usually ignored because little data on the time history are available.
Development of time and space variant reservoir models is underway, and will help refine marine radiocarbon dating (see Goldfinger et al., 2009). Because the sedimentation rates in the deep sea are relatively stable over periods of interest to paleoseismology, radiocarbon ages can be corrected for such factors as basal erosion and sample thickness. To correct ages for the thickness of the radiocarbon sample, it is necessary to subtract the time representing half the sample thickness from the $^{14}$C age. This correction attempts to bring the age as close as possible to the age of the deposition of the turbidite (barring basal erosion). Sedimentation rate curves can be constructed for each core using the pelagic or hemipelagic interval thicknesses and radiocarbon data, and this simple correction calculated from the curves.
**Figure 2B.7:** (A) Detailed image of a Cascadia paleoseismic turbidite, its subunits and preferred $^{14}$C sampling site. In this example, the turbidite tail/hemipelagic boundary is distinct visually, and variably disturbed by bioturbation. While turbidite bases can be erosive, dating is commonly done
Basal erosion is a primary concern when dating landslides or turbidites below their bases. In the case of landslide deposits, it may be advantageous to date material directly above the deposit as it is both easier to sample and lacks the erosion concern. One can mitigate the basal erosion problem to some extent by estimating the degree of basal erosion. Examination based on the morphology of the turbidite bases for the degree of roughness and obvious truncations is of value, though such analyses are imperfect. Another technique is to examine the underlying hemipelagic sediment for thickness variations among a local group of cores with the assumption that erosion is likely the primary cause of such variability. Missing section can be estimated from the difference between the thickest and the other intervals (Goldfinger et al., 2008, 2009; Gutierrez-Pastor, in press). This method obviously underestimates erosion in the case where all samples are eroded.

Radiocarbon ages can be reported in a variety of ways, and these vary considerably. These are sometimes reported as simply a 1σ or 2σ range, a conservative approach. However, the probability distribution generated during the calibration process contains more information, including a probability peak that may also be significant. The probability peak can be considered the most likely age of the event of interest (Blaauw et al., 2005); though of course many other external factors are involved in the correspondence between the 14C age and the event age. Goldfinger et al. (2007) show the strong tendency for the probability peaks to lie along the age model sedimentation rate, despite smoothing of the rate curve. Goldfinger et al. (2009) suggest the prudent use of probability peaks and ranges as shown by Goslar and others (2005).

To develop an age model for sediment cores, it is necessary to determine hemipelagic thickness between events in the case of multiple turbidites or landslides. To establish this thickness, the boundary between the gradational turbidite tail and the overlying hemipelagic sediment must be determined as precisely as possible, though it is uniquely difficult. The reason is that the differences between the very fine-grained turbidite tail and the overlying hemipelagic may be nearly nonexistent. In the case of Cascadia cores, many tend to have obvious boundaries that are clearly visible to the eye (Goldfinger et al., 2008, 2009). For many other regions, the problem is more difficult. Many attempts have been made to find universal

from planktonic forams in the upper part of the underlying hemipelagic interval as the least problematic option. Typical sample location shown, with small “gap” above the sample. (B) Detail from Core RR0507-25TC event T4 along the Northern San Andreas margin. Example grain size analysis, magnetic susceptibility/density signatures and X-radiography in turbidites T19 and T20 in core 24GC below the Gualala–Noyo–Viscaino channel confluence. Light tones in the X-radiograph represent dense sand/silt intervals; darker gray tones represent clay/mud. Oval dots are grain size samples. Heavy trace is the magnetic susceptibility signature. Right plot is percent sand (obtained with Coulter laser counter method). The good correspondence between grain size, density, and magnetic susceptibility for the lithologies in both Cascadia and NSAF cores is established with selected analyses and permit the use of density and magnetics as mass/grain size proxies that show much greater resolution than possible with grain size analysis. These typical turbidites are composed of 1–3 fining upward sequences, each truncated by the overlying pulse. No hemipelagic exists between pulses, indicating the three pulses were deposited in a short time interval. Only the last pulse has a fine tail, indicating final waning of the turbidity current. We interpret these signatures as resulting from a single multipulse turbidity current. Number of coarse pulses commonly remains constant in multiple channel systems for a given event. Source provenance affinity for each sand pulse is shown to the right. Mineralogically distinct sandy units stacked vertically in order of arrival at a confluence near the core site. See Goldfinger et al. (2007) for further details and core locations.

Modified after Goldfinger et al. (2007, 2008). (See Color Insert.)
methods for defining this boundary including clay fabric orientation (O’Brien et al., 1980; Azmon, 1981), color (Rogerson et al., 2006), hydraulic sorting of microfossils (e.g., Brunner and Ledbetter, 1987), XRF and XRD (e.g., Bernd et al., 2002), and grain size (the most common method, i.e., Brunner and Ledbetter, 1987; Joseph et al., 1998; St.-Onge et al., 2004), resistivity, and other methods. Once an acceptable criterion for the boundary is determined, sedimentation rates and an age model can be constructed.

2B.2.2.2 OxCal Analysis

OxCal is radiocarbon calibration software that also includes multiple methods to allow the use of external age constraints, multiple $^{14}$C ages and geological constraints such as sedimentation rates to constrain radiocarbon ages. The technique uses Bayesian statistics to combine multiple probability distributions and trim probability density functions (PDFs) (Ramsey, 1995, 2001). The external constraints may include (1) the time represented by sediment deposited between events, (2) historical information, (3) stratigraphic ordering, and (4) other external stratigraphic constraints such as dated ashes, pollen, or other biostratigraphic markers (Biasi et al., 2002; Goldfinger et al., 2007). Where age data are missing, sedimentation rates alone can be used to model event ages. Since calculated sedimentation rates are also dependent on the radiocarbon ages, and on basal erosion, there is some unavoidable circularity in this process unless varves or other independent rates are available (e.g., Kelsey et al., 2005). Figure 2B.8 shows an example of this using hemipelagic sedimentation and historical constraints for the AD 1700 Cascadia earthquake and the NSAF 1906 and penultimate NSAF earthquakes. Further details of constrained age models are give in Goldfinger et al. (2007, 2008).

2B.2.2.3 Sedimentation Rate Ages

The age model of a marine core with good age control is a powerful tool. Using the sedimentation rates and hemipelagic thicknesses, one can calculate the age of an undated event based on a dated turbidite below or above (or both when possible) using sedimentation rates alone. Goldfinger et al. (2009) present calculations for undated cascadia events, while, Gutierrez-Pastor et al. (in press) present additional analysis of hemipelagic age calculations from the Cascadia margin with hemipelagic intervals treated as a semi-independent event-time line. A similar analysis is presented in Kelsey et al. (2005) for a tsunami record in Bradley Lake, Cascadia margin.

2B.2.2.4 Event Ages and Potential Biases

The question of how well the radiocarbon ages from marine deposits represent earthquake ages is complex. In land paleoseismology, ages commonly represent maximum or minimum ages when dated using sample material below or above the event, respectively. Typically the best one can do is to collect material from as close below and as close above an event, and refer to these ages as “close maximum” and “close minimum” ages, respectively (i.e., Nelson et al., 2006, 2008). The sample materials are commonly detrital, and thus certain to be of different age than the earthquake (see Chapter 2). These are commonly reported in the literature and usually indicated on space–time diagrams with arrows pointing upward or downward for maximum and minimum ages, respectively. Marine ages may, however, include reasonable attempts to correct known biases based on continuous marine sedimentation. The sedimentation rate corrections, erosion analyses, and OxCal analyses using hemipelagic intervals discussed previously are designed to approach event ages by attempting to remove these biases. These tools are unavailable in most land settings due to the absence of continuous sedimentation, though are used when other constraints are available (see Kelsey et al., 2005). Goldfinger et al. (2009) report several examples of testing these methods against events of known age, with good results (see also Figure 2B.8).
Sequence N SAF 1906 test

1. Calibrate the RC date for the upper-most event: 1913 (1898-1940)
2. Date of coring (1999) minus 'H1' (=83.6): 1916
3. Age of calibrated penultimate event (1724(1647-1819)) plus 'H2' (=137): 1861 (1784-1956)
4. Use the Sequence option in OxCal with all available data (preferred option):

   - **simple calibration** (option 1)
   - **OxCal (constrained)** calibration (option 4)

**Marine data from Hughen et al (2004); Delta_R 341±33; OxCal v3.10 Bronk Ramsey (2005); cub r:5 sd:12 prob usp[chron]**

Sequence Cascadia 1700 test

- **T1 134.1%**
- **Gap 299**
- **C_Date Date of Collection 100.0%**
- **Boundary**
- **AD 1700 calendar age from Japanese records**

**Calendar date**

- 1500BP
- 1000BP
- 500BP
- 0BP

**Marine data from Hughen et al (2004); Delta_R 341±33; OxCal v3.10 Bronk Ramsey (2005); cub r:5 sd:12 prob usp[chron]**

Figure 2B.8: OxCal methods example using the well constrained 1906 San Andreas earthquake and the AD 1700 Cascadia event. The left panel shows the hemipelagic (H) data determined from visual observation, physical property data, smear slide mineralogy and X-radiography. H data are then input to OxCal with raw 14C ages converted to time via sedimentation rate curves developed for each site. Right panel shows four ways to calculate the age of the 1906 earthquake, with the preferred method being the use of underlying and overlying hemipelagic intervals, historical data (no written record of an earthquake between the date of the first San Francisco Mission built in 1769 and the 1838 earthquake). Lower panel shows result of similar application of constraining hemipelagic data to the AD 1700 Cascadia earthquake, the age of which is known independently (Satake et al., 2003). This method commonly resolves the ambiguities inherent in radiocarbon dating where probability density functions (PDFs) have multiple peaks or broad distributions due to the slope or complexity.
2B.2.2.5 $^{210}$Pb and $^{137}$Cs Activity

The $^{210}$Pb method can be used to constrain the ages of young events and also to determine whether or not piston cores captured the youngest material at the seafloor. The free-falling piston corer sometimes does not sample the interface, which is blown away by the force of the falling core barrel, whereas the slowly lowered trigger core almost always includes the seafloor. $^{210}$Pb activity rates can be used to either determine the age of the uppermost sediment, or determine that the uppermost material was older than the maximum typical age when $^{210}$Pb reached background levels (~150 years; Robbins and Edgington, 1975).

Numerous observations from multicore samples and submersibles show that there is a very low-density material near the nepheloid layer at the seafloor. This layer is usually not recovered in piston and gravity cores, but is easily observed in multicore samples. The logarithmic decay of $^{210}$Pb begins below the mixed layer, and most $^{210}$Pb analyses assume that the mixed layer is entrained in any turbid flow and completely removed from the record we observe in sediment cores. This apparently presents no significant problem, as there is no time lost by removing the mixed layer as its $^{210}$Pb age is constant and near zero on the seafloor prior to the rapid deposition of turbidites and landslides (Nittrouer, 1978).

$^{137}$Cs is a similar technique for even younger materials. The half life of $^{137}$Cs is 30.3 years. Its presence is due to the atmospheric testing of nuclear devices during the 1950s and early 1960s (Schuller et al., 1993, 2002). Since that time, there has been no $^{137}$Cs released to the atmosphere following the termination of atmospheric nuclear testing. Very recent small releases include the Chernobyl accident and testing by India and Pakistan. When using $^{137}$Cs activity in sediments, most investigators assume that the maximum value is associated with the high fluxes of $^{137}$Cs between 1962 and 1965, with the peak value commonly assumed to be 1963 ± 2 years, giving a sharp peak that can be used to constrain sedimentation rates and calculate ages of postbomb strata. In some settings including marsh peat and organic materials, $^{137}$Cs has been shown to be mobile and not useful for dating. The technique works well in clay rich systems.

2B.2.2.6 Bioturbation and its Effect on Radiocarbon Dating of Interseismic Hemipelagic Sediments

A number of attempts to test the dependence of vertical rates of mixing during bioturbation on various parameters have been made. When considering single species, it has been shown that rates can be dependant on temperature, particle size, and particle shape (Wheatcroft et al., 1992 and references therein). Abyssal plain temperatures are relatively constant, and therefore unlikely to contribute to variable bioturbation. Thus we would like to know what effect particle size has on vertical mixing rates to evaluate radiocarbon age results in terms of foram size within samples, as well as any effect such rate changes may have toward biasing $^{14}$C ages in either direction.

The process of bioturbation is highly complex, and equally complex to unravel. Experimental results from several settings suggest, however, that bioturbation in the deep sea is dominated by deposit feeders, and that deposit feeders in turn preferentially ingest and retain fine particles (Thomson et al., 1988, 1994; Wheatcroft, 1992). The impact of this in the context of dating marine deposits is that relatively large particles such as foraminifers used for dating are not selected by deposit feeders for retention, and...
apparently not vertically mixed as much as the finer fractions of material. For dating of turbidites, these are important results, and may help explain the surprising consistency we see in dating correlative turbidites when other variables such as reservoir age, basal erosion, and contamination are minimized.

2B.2.2.7 Stratigraphic Datum Ages

In Cascadia, the widespread deposition of ash sourced from the eruption of Mount Mazama (now Crater Lake) provides a clear datum throughout most of the Cascadia Basin system and provides independent age control. The age of the cataclysmic Mazama eruption is well constrained by recent work, yielding an age of 7630 ± 150 cal BP (Zdanowicz et al., 1999) from Greenland ice cores, and 7600 ± 29 in British Columbia lake sediments (calibrated from Hallett et al., 1997). Throughout Cascadia Basin, the first turbidite containing the Mazama ash is easily identified and has been dated in five localities, with an average age of 7130 ± 120, ~500 years after the Mazama eruption. Earlier work identified this Mazama ash-bearing turbidite as the thirteenth event down from the surface in many Cascadia basin cores (Adams, 1990; Goldfinger et al., 2003a,b). Subsequent work demonstrated that Rogue Channel events could be well correlated locally, as well as correlated to other Cascadia Basin sites (Goldfinger et al., 2008; see below) and that the Mazama ash first appearance was in the fourteenth margin-wide turbidite down from the surface, not the thirteenth as in other systems. Similar stratigraphic control can be found in many other settings, such as the Sumatran margin where multiple tephras with good elemental fingerprints constrain the turbidite sequence there (M. Salisbury, Oregon State University, personal communication 2008). Similarly, turbidite ages have been constrained by tephra markers in Lake Biwa, Japan (Inouchi et al., 1996). Twenty turbidites were correlated to the historical record of earthquakes, and their ages calculated using sedimentation rates above the known tephra markers. Similarly, Noda et al. (2008) used a tephra marker to help support the dating and correlation of turbidite stratigraphy on a fan setting in the Kurile Trench.

2B.3 Locating Primary Evidence: Active Faulting and Structures

2B.3.1 Direct Fault Investigations

2B.3.1.1 Wecoma Fault, Cascadia Subduction Zone

An early example of slip rate determination based on submarine imaging and limited sampling is a group of unusual active strike-slip faults cutting both plates of the Cascadia subduction margin (Appelgate et al., 1992; Goldfinger et al., 1992, 1996, 1996b, 1997). Investigation of one of these, the Wecoma Fault revealed that a youthful submarine channel was clearly offset in high-resolution deep-towed sidescan sonar imagery (Figure 2B.9; Appelgate et al., 1992; Goldfinger et al., 1992). Cores showed that the channel had been abandoned at the start of the deglacial, setting a timeframe for the measured offset and establishing a minimum slip rate of 8.5 ± 4 mm/yr. No information on individual slip events could be determined from these data.

Matching of subsurface sedimentary packages imaged in a detailed seismic reflection grid allowed estimates of total fault slip since inception of the Wecoma Fault, and linkage of a key reflector to DSDP site 174 allowed calculation of the average slip rate since fault inception of 7–11 mm/yr, similar to the late Quaternary rate. Four of the nine mapped faults were investigated with the Alvin and SeaCliff deep diving submersibles, and two by the shallow Delta submersible. These faults (the Wecoma, Thompson Ridge Daisy Bank, and North Nitinat Faults) all exhibited evidence of recent surface rupture.
At the intersection of the Wecoma Fault, with the frontal thrust ridge of the accretionary wedge, samples and video showing strongly developed sub-horizontal slickensides were collected (Goldfinger et al., 1997). Samples revealed carbonate chemistry elevated in $^{3}$He, indicative of rupture of the Juan de Fuca plate slab, also indicated in reflection profiles (Sample et al., 1993).

Figure 2B.9: (A) Interpreted map view of the Wecoma Fault, central Oregon USA, based on sidescan imagery, deep reflection profile, shallow sub-bottom profiles, and ALVIN dives. The main fault bifurcates as it intersects the megathrust, labeled “frontal thrust” creating a triangular pop-up. Two thrust ridges cut and offset by the basement fault shown, with sigmoidally deformed anticlinal axes. (B) Schematic view of retrodeformed trench-fill wedges used to establish net slip since fault inception. (C) Composite 3D block diagram of the intersection of the Wecoma Fault with the Cascadia megathrust, viewed toward the northeast. Migrated seismic sections scaled in two-way travel time (seconds). AP, abyssal plain section; AF, Astoria Fan; A, motion away; T, motion toward; SV, seaward vergent thrust, LV, landward vergent thrust; OC, Oceanic Crust. Modified after Goldfinger et al. (1992, 1997).
The North Nitinat Fault also offset a seafloor channel and localized an elongate mud volcano aligned along the fault (Goldfinger et al., 1997). Shallow water observations of the Daisy Bank Fault on the Oregon shelf revealed a fresh scarp of ~0.5 m in height, offsetting the Holocene–Pleistocene conformable horizon, and breaking carbonate pavements formed during a previous interseismic period, indicating high accelerations (Goldfinger et al., 1996).

### 2B.3.1.2 Lake Tahoe, California

In cases where faults traverse the shoreline, both onshore and offshore techniques can be used to best advantage. Offshore, geophysical imaging is efficient, can be of very high resolution, and is commonly relatively free of cultural noise that can plague onshore high-resolution geophysics. Fault morphology is commonly well expressed in the marine environment where erosion is minimized and scarps preserved longer due to the reduced gravitation. This makes bathymetric, sidescan sonar, and reflection imaging of submarine faults unparalleled except perhaps for faults in desert environments. Many of the same cross cutting relationships such as landsliding, offset channels, offset structures, and other features are available in the submarine environment. On the other hand, sampling and dating individual events is more difficult and costly offshore, but relatively inexpensive onshore with trenching and coring. An excellent example of the use of both techniques has been developed for faults in the Lake Tahoe Basin.

Kent et al. (2005) report evidence of deformation across three major fault strands within the Lake Tahoe Basin based on combination of high-resolution CHIRP seismic, a bathymetric grid combining airborne laser and multibeam sonar, and sediment cores (Figures 2B.4 and 2B.10). These faults offset submerged erosional terraces of late Pleistocene age (19.2 ± 1.8 ka) and record 10–15 m of vertical deformation. A major submarine landslide, the McKinney Bay slide, has spread blocks across much of the central lake floor. This deposit is also offset vertically across the Stateline fault by ~21–25 m.

Age constraints on the landslide deposit, and thus on the fault slip rates are uncertain, and reported variously as 60 ka (Kent et al., 2005), 15–17 ka (Moore et al., 2006) 300 ka (Gardner et al., 2000), to Holocene (Schweickert et al., 2000). Kent et al. (2005) calculate a deformation across several marker beds, and ^14^C and optically stimulated luminescence (OSL) age control suggest an extension rate across the Lake Tahoe basin that of 0.4–0.5 mm/yr, assuming the 60 ka age of the McKinney Bay Slide. Moore et al. (2006) have suggested this slide generated a large tsunami, which may have had a role in the deposition or modification of linear boulder lines on the western shelf of the lake Seitz and Kent (2005) investigated one of the Lake Tahoe faults, the Incline Village Fault using a grid of high-resolution CHIRP seismic profiles offshore imaging the fault zone in considerable detail, and suggesting multiple events offsetting the mostly glacial stratigraphy and multiple colluvial wedges. Onshore, the fault scarp was trenched with a 7.5 m deep trench across the 5-m-high scarp, yielding event ages of 500 years, ~36.7 ka, and an older as yet undated deposit, demonstrating the utility of the integrated onshore–offshore technique (Seitz and Kent, 2005) (Further details of the most recent earth quake on one of the Tahoe faults are given in Brothers et al., 2009).

### 2B.3.1.3 Palos Verdes Fault

Offshore Los Angeles, the Palos Verdes Fault, represents one of many proximal seismic hazards to the Los Angeles area represented by the San Andreas–parallel Peninsular Range faults. The Palos Verdes Fault is on the shallow shelf for the most part and partially located onshore on the Palos Verdes Peninsula (Figure 2B.11). Because the onshore part of the Fault is located in a heavily populated and industrial area, onshore investigation is all but impossible. Offshore, McNeilan et al. (1996) used high-resolution seismic-reflection profiles and borehole data from the Los Angeles Outer Harbor to estimate the slip rate for the San Pedro segment of the Palos Verdes Fault based on subsurface piercing point offsets. Seismic profiling...
revealed two paleochannels offset by the fault, as well as mismatching structure contours. Using numerous $^{14}$C ages, they then estimated the slip rate to be between 2.7 and 3.0 mm/yr for the past 7.8–8.0 ky based on reconstruction of offset paleochannels and subtle structure contours (Figure 2B.11). The slip rate obtained in this way is in good agreement with estimates based on uplift of marine terraces onshore (Ward and Valensise, 1994) using the combined vertical and horizontal components of separation.

This study is a classic application of land paleoseismic techniques adapted and applied to a submarine fault. Seismic profiles serve as geophysical trenches, and can be collected in a much larger spatial array and longer extent. While this technique lacks the spatial resolution to identify individual event details as is common in a trench wall, the slip rate is probably more robust due to the larger extent of sampling of the fault trace. McNeilan et al. (1996) used their reconstructed slip rate and observed segmentation to calculate that the fault is capable of generating an $M_w = 7.0–7.2$ earthquake every 400–900 years.

Figure 2B.10: (A) Deep-towed Edgetech CHIRP reflection profile of across the Stateline Fault (SLF), Lake Tahoe California. Top of the McKinney Bay slide complex is offset 21 m, with possible small colluvial wedges visible against the fault plane. Location map shown in B. From Kent et al. (2005). Reprinted with permission from the Geological Society of America. (See Color Insert.)
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Figure 2B.11: Example of subsurface mapping of piercing point offsets in the subsurface along the Palos Verdes Fault, southern California. (A) Map of the Los Angeles Outer Harbor showing geologic exposures, Holocene paleochannel Holocene fault strands, and zone of Pleistocene fault deformation (hachured). For the northern paleochannel only, the thalweg and limits of the sub-channel are shown. (B) Map of piercing lines of northern 7.8–8 ka paleochannel across the Palos Verdes Fault; solid lines represent thalwegs and dashed lines represent other piercing lines. From McNeilan et al. (1996). Reprinted with permission from the American Geophysical Union.

Marlow et al. (2000) used high-resolution multibeam sonar to map the surface trace of the fault extending to the southwest of the McNeilan study, and showed strong evidence of surface faulting, associated active anticlines, and interaction with the sub-parallel Avalon Knoll Fault farther offshore. Fisher et al. (2004) and Bohannon et al. (2004) used deeper seismic reflection profiles to characterize the geometry of the Palos Verdes Fault, and its along-strike variability in dip, to link the structure to growth of anticlines such as Lasuen Knoll along the fault trace (Figure 2B.12).

2B.3.1.4 San Clemente Fault, California

The southern California Borderland province hosts many other active marine structures that have been the targets of paleoseismic and active fault studies. The San Clemente Fault is one of the most significant of
Miocene? rocks at shallow depth or truncated at the seafloor

Active San Gabriel Channel
San Gabriel Channel
Main strand of the Palos Verdes Fault

Seafloor Ridge "B" (Plate 1)
Seafloor Ridge "A" (Plate 1)

Transverse scarp
Sharp scarp (Figure 10)

Underlying rocks extensionally deformed

San Pedro Shelf

Lasuen Knoll

Central Part Line 068

Holocene to late Pleistocene interval
San Gabriel Canyon

Unnamed Fault

Lasuen Knoll (north end)
Palos Verdes Fault Zone

Vertical Scale (500 m)
Horizontal Scale (500 m)

Twtt (s)
CDP

Basement interval

WBM

Holocene to late Pleistocene interval
San Gabriel Canyon

Unnamed Fault

Lasuen Knoll (north end)
Palos Verdes Fault Zone

Vertical Scale (500 m)
Horizontal Scale (500 m)

Twtt (s)
CDP

Basement interval

WBM
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the Peninsular Range Faults, and traverses the borderland from south of the US–Mexico border to its intersection with the limit of the province at the Channel Islands Thrust (CIT), the southern limit of the Western Transverse Ranges (Figure 2B.13).

The San Clemente Fault is an active structure with instrumental seismicity and occasional moderate dextral earthquakes along its mapped trace. The San Clemente Fault has apparently been a strike-slip fault for all of its history, whereas many of the peninsular range faults have an earlier extensional history derived from the failed rifting of the Borderland. The San Clemente Fault clearly offsets large scale features in the borderland that can be retrodeformed in a relatively straightforward way. Construction of a hybrid multibeam/singlebeam bathymetric grid for the borderland revealed both broad and fine scale tectonic geomorphic relationships along the San Clemente, San Diego Trough and other fault systems. Slip sense for the San Clemente Fault is revealed by offset drainages, offset basement highs, and the numerous restraining and releasing bends that control the vertical tectonics on both fine and regional scales. Retrodeformation of regional piercing points along the San Clement Fault, including a dextral separation of San Clemente Island from Fortymile Bank to the southeast established the regional net slip and average slip rate since inception of the fault of ~50–62 km and ~7 mm/yr. On a smaller scale, numerous restraining and releasing bends control the development of related folds along the San Clemente fault, also indicated by shifting channels and Holocene–Pleistocene growth strata (Goldfinger et al., 2000; Legg et al., 2007). Superimposed on this broad uplift are four smaller restraining-releasing bend pairs, mirroring the larger uplift that results from a left bend in the main fault trace near the Descanso Basin. ALVIN observations of the San Clemente fault on the northern flank of Navy Fan (Legg et al., 2007) reveal a recent Holocene scarp 0.3–1.5 m in height with apparent horizontal slickensides exposed. The scarp is interpreted as a single event scarp, indicated by the lack of multiple slope breaks, and uniform “weathering” and bioturbation of the exposed Holocene and late Pleistocene strata. Scarp height suggests a Holocene event greater than Mw = 6 (Goldfinger et al., 2000).

2B.3.1.5 Marmara Sea

The Sea of Marmara in Turkey represents a series of linked pull-apart basins along the North Anatolian Fault (NAF). The 1999 Izmit earthquake, coupled with the likelihood that the next NAF earthquake may strike Istanbul, has spurred research efforts in the Sea of Marmara, which contains the submarine segment of the NAF and the likely site of the next significant earthquake. Le Pichon et al. (2001, 2003) and Armijo et al. (2002) present results from high-resolution multibeam bathymetric surveys with backscatter data, sidescan sonar, and seismic reflection profiles newly acquired in the Marmara basin. These data revealed a series of en echelon pull-apart basins within the larger Marmara basin, interpreted by Armijo et al. (2002) as the apparent lack of a throughgoing dextral NAF. LePichon et al. (2003) argue for a main Marmara fault that ruptures most of the system, and a Çınarcık basin segment to the east with extension expected there based on GPS data and kinematic reconstruction. They compare this proposed rupture mode to

Figure 2B.12 (A) Shaded relief view of high-resolution multibeam bathymetry along the Palos Verdes Fault showing tectonic features of this strike-slip fault. Location of Figure 2B.11 shown on shelf at upper portion of image. Other figure callouts refer to the original publication. From Fisher et al. (2004). Reprinted with permission from the Geological Society of America. (B) Migrated multichannel reflection profile across the Palos Verdes Fault and Lasuen Knoll, a small restraining bend uplift along the fault (Legg et al., 2007). Line of section A–A’ shown in A above. Figure from Bohannon et al. (2004). (See Color Insert.)
Figure 2B.13: (A) Bathymetric compilation map of the southern California Borderland showing the San Clemente Fault and the San Diego Trough Fault (SDTF). (B) Photograph from the DSV Alvin of fault scarp along the San Clemente Fault crossing the Navy fan (location in B). Sub-horizontal lineations may be slickensides consistent with strike slip motion. The scarp is composed of mud and layers of shells associated with ancient benthic communities at former cold seeps. Holocene scarp 0.3–1.5 m in height appears to be a single event scarp, indicated by the lack of multiple slope breaks.
historical earthquakes to support a correspondence between onshore damage and these two primary segments. Rangin et al. (2004) further propose that the throughgoing fault is a late development of the most recent 100–200 ka, and that this recent propagation of the throughgoing main Marmara fault deactivated the pull-apart basins, as evidenced by undisturbed sedimentary overlap of the bounding transfer faults, inversion of the pull-apart basins, and the apparent crosscutting of the main Marmara fault across the corners of the en echelon pull aparts (Figure 2B.14). In this model, the preexisting pull-apart basin system accumulated about 30 km of dextral slip and probably appeared in late Pliocene or early Pleistocene time. Subsequently, the Sea of Marmara has been the site of progressive localization of strain. This model bears similarities to experimental model results of evolving strike-slip systems. Thus in a relatively short time since the 1999 earthquake, collection of marine geophysical data has resulted in the rapid evolution of a series of tectonic models for the area that was previously quite poorly known. Discussion of paleoseismic stratigraphy in the Marmara Sea in relation to historical earthquakes is included in a subsequent section.

2B.3.2 Off-Fault Investigation

2B.3.2.1 Vertical Tectonics in a Strike-Slip Setting: Channel Islands Thrust and the Catalina Ridge–San Clemente Fault Zone

Bathymetric data can be used in a number of ways and are particularly amenable to strike-slip environments as previously discussed. Another way to use these types of data is to establish strain markers such as previously level low-stand shorelines as strain markers. Chaytor et al. (2008) used submerged last glacial maximum (LGM) and younger paleoshorelines preserved around the Northern Channel Islands submarine banks atop the Santa Cruz–Catalina Ridge to determine the vertical strain history at the intersection of the Santa Cruz–Catalina Fault and the southern Transverse Ranges, marked by the CIT. They used high-resolution multibeam mapping combined with submersible observations to establish the nature of the shorelines (Figure 2B.15). The morphology of the slope breaks and surficial sediments revealed clear evidence of the former wave-cut shoreline angles, with a sharp slope break, wave-cut undercuts, coarse sediment on the former shoreface, decreasing in grain size downslope of the gently sloping planar platform. Intertidal mussels and barnacles were collected manually from the submersible along the submerged shorelines, presently at a depth of 100–130 m. Radiocarbon results bracket the LGM (LGM ~ 120 m water depth) at ~19 ka. On the eastern Northern Channel Islands platform, as much as and uniform "weathering" and bioturbation. The lightly bioturbated fresh scarp offsets Holocene and late Pleistocene strata, indicating a Holocene event that likely had a magnitude greater than 6 (Goldfinger et al., 2000). Photo by C. Goldfinger, from Legg et al. (2007). (C) View of a restraining bend along the San Clemente Fault south of Navy Fan showing multiple pull apart and restraining bend features, superimposed on the larger uplift which itself is a restraining bend uplift due to a 5° strike change in the San Clemente Fault visible in (A). Location shown in (A). Channel at left is presently on the flank of the uplift, reflecting recent growth of this feature. (D) Retrodeformed San Clemente and SDTFs using morphologic and geologic piercing lines (Goldfinger et al., 2000). San Clemente Fault (SCF) has a minimum horizontal separation of 50 km based on four piercing points (two are shown). The SDTF horizontal separation is 32 km, with 15 km extension as well (extension retrodeformation partially shown here to illustrate fit of offset features. (A and C) Reprinted with permission of the Royal Society of London. (See Color Insert.)
Figure 2B.14: (A) Detailed morpho-bathymetry of the Marmara Sea's Central Basin, location of giant piston cores and seismic profiles. (B) 3.5 kHz profiles across the Central Basin, showing active faults and evidence for a "homogenite." (C) Close-ups of selected portions of core MD01-2431, (a: conjugate microfractures; b: microfracturing with possible sealing by coeval turbidite arrival; c: possible in situ liquefaction, evidenced by ball-and-pillow –b–a–p– structure; d–g: details of the pre-Late Glacial event. In the continuous core section from 10.50 to 12.00 m X-ray scanning shows a constant orientation of microfractures). From Beck et al. (2007). Reprinted with permission from Elsevier. (See Color Insert.)
1.50 ± 0.59 mm/yr of late Pleistocene to Holocene uplift of the islands above the blind CIT was observed based on the uplifted shoreline. This result is higher than onshore terrace uplift estimates over a period of 125 ka since stage 5e terrace formation (Pinter et al., 1998).

South of the intersection of this fault with the CIT, similar shorelines were observed rimming the submarine Pilgrim Banks atop the Santa Cruz–Catalina Ridge. Shorelines there show no net vertical tectonic motion, but are instead tilted to the north, possibly reflecting flexural bending or limited underthrusting of this block beneath the CIT.

Collectively, the submerged shorelines revealed significant uplift from underthrusting along the CIT, and northward tilt of the underthrust block. It also appears that a significant amount of differential motion at the intersection of the Peninsular Range faults and the southern Transverse Range thrusts offshore may be distributed into upper-crustal deformation both at and south of the intersection, along the length of the major dextral fault systems. Detailed bathymetry and reflection profiles suggest some of this deformation is partitioned into splay fault terminations and thrusting on the western side of the Santa Cruz–Catalina Ridge, and more subdued deformation in the basin to the east. This may be an effect of the “subduction” of the dextral strike-slip fault beneath the CIT, and the resulting change in slip rates on the CIT across the intersection point with the strike-slip fault. Similar kinematics was reported for subduction of strike-slip faults along the Cascadia margin, discussed previously (Goldfinger et al., 1997).

Other notable localities in which detailed submarine faulting studies of individual faults have been carried out, though not to the level of detail in these case histories, include the Gulf of Corinth, the Carboneras Fault off southern Spain, offshore Lebanon, and many subduction zones including Cascadia, Nankai, Hikurangi, Sumatra Costa Rica, Chile, and others.

2B.4 Locating Secondary Evidence: Landslides, Turbidites, Submarine Tsunami Deposits

The classic paper by Heezen and Ewing (1952) demonstrated that large offshore earthquakes can trigger turbidity currents having regional extent. They described the Grand Banks turbidity current, which was triggered in the epicentral area of a magnitude 7.2 earthquake on 28 November 1929. This event involved detachment and downslope movement of submarine sediment along 240 km of the continental shelf; after traveling 650 km from its source, the turbidity current still was moving faster than 20 km/h and therefore probably continued for hundreds of kilometers. Heezen and Ewing (1952) postulated that the earthquake triggered submarine slumps along an extensive length of the continental shelf corresponding to the epicentral zone of the earthquake and that these slumps transformed into turbidity currents that moved as rapidly as 100 km/h down slopes averaging only about 1.5°.

The triggering of turbidity currents and landslides from submarine canyons, shelf edges, and seamount edifices are becoming reasonably well known. In particular, turbidity currents triggered along the Cascadia margin, the northern San Andreas margin, Chile, the Japan Trench and other localities are under investigation and yielding coherent earthquake records. Many of these events have been linked temporally to onshore tsunami deposits and are becoming recognized as viable event pairs that document the occurrence of earthquake-triggered tsunamis. These events can be dated and correlated in the marine environment, providing long continuous records that also provide good evidence for spatial continuity.

The use of secondary evidence such as landslides and turbidites adds some complexity to this aspect of paleoseismology. The techniques do not use fault outcrops because the faults are inaccessible and must
Figure 2B.15: (A) Shaded relief map of bathymetry compilation of the southern California Borderlands showing two study sites at Santa Cruz Island and Pilgrim Bank. (B) View of the bathymetry, NCI-S1 paleoshoreline, and terrace features on the southern Northern Channel Islands platform shelf edge between Santa Cruz and Anacapa Islands. The trace of the Santa Cruz Island fault (SCIF) is indicated as it crosses onto the platform, where it is likely related to destabilization of the slope. Inset: profile A–A’, showing the morphology of the LGM terrace, with approximate locations of the SCIF and an additional fault indicated (fault movement indicated by X-away, O-toward). (C) Composite schematic diagram of paleoshoreline features observed during submersible dives on the Northern Channel Islands platform and Pilgrim Banks. Examples of these features can be seen in photographs taken from the submersible: (A) Large, well-preserved Mytilus californianus shells on bench, Pilgrim Banks. (B) Notched, undercut rock outcrop on south side of Santa Cruz Island. (C) Rounded cobbles–boulders on probable paleoshoreline between Santa Cruz and Anacapa Islands, southern Northern Channel Islands platform. (D) Map of paleoshorelines and benches–terraces on Pilgrim Banks, based on analysis of the bathymetry and submersible observations. Terrace–bench areas are shown in white. Possible pre-LGM inner edges–terraces are indicated. Line of profile A–A’ is...
demonstrate that the events they are investigating are uniquely generated by earthquakes and not some other natural phenomenon. Nevertheless, these problems can be overcome, and the techniques can be powerful tools for deciphering the earthquake history along an active continental margin. These methods are complementary: the onshore record can provide temporal precision for the most recent events via radiocarbon dating, coral chronology and dendrochronology (tree-ring dating), while the marine sedimentary record generally extends further back in time, more than enough to encompass many earthquake cycles. In recent years, turbidite paleoseismology has been attempted in Cascadia (Adams, 1990; Goldfinger and Nelson, 1999; Blais-Stevens and Clogue, 2001; Goldfinger et al., 2003a,b, 2007, 2008, in review), Puget Sound (Karlin and Abella, 1992; Karlin et al., 2004), Japan (Inouchi et al., 1996), the Mediterranean (Kastens, 1984; Anastasakis and Piper, 1991; Nelson et al., 1995), the Dead Sea (Niemi and Ben-Avraham, 1994), northern California (Field et al., 1982; Field, 1984; Garfield et al., 1994; Goldfinger et al., 2007, 2008) Lake Lucerne (Schnellmann et al., 2002), Taiwan (Huh et al., 2006), the southwest Iberian margin (Viscaino et al., submitted), the Chile margin (Blumberg et al., 2008, 2009, in review), the Marmara Sea (McHugh et al., 2006), the Sunda margin (Patton et al., 2007) and even the Arctic ocean (Grantz et al., 1996). Results from these studies suggest the turbidite paleoseismologic technique is evolving as a useful tool for seismotectonics.

2B.4.1 Distinguishing Earthquake and Nonearthquake Triggering Mechanisms

In off-fault paleoseismology, considerable effort must go toward distinguishing earthquake and nonearthquake sources. In the following sections, we consider this issue in some detail.

Common stratigraphic evidence of earthquakes includes submarine landslides and turbidity currents. Triggering events for these deposits may include (1) earthquakes; (2) volcanic explosions; (3) tsunami; (4) subaerial landslides into the marine environment; (5) storm wave loading, and (6) hyperpycnal flow. These primary triggers are distinguished from factors that may destabilize slope through longer term processes, such as sediment self-loading, gas hydrate thermal destabilization, sea-level change, shelf edges destabilized by groundwater input, volcanic seamount or island edifice destabilization, tectonic folding/tilting, and other factors. Triggering mechanisms have been discussed by Adams (1990), Nakajima (2000), Goldfinger et al. (2003a,b, 2008, 2009). Factors such as gas hydrate destabilization, sea-level change, tectonic steepening, and so on are factors that reduce seafloor stability, but do not generally trigger submarine mass movements. For example, the Storegga slide generated a large tsunami and occurred as a result of the massive deposition of glacial sediments and associated gas hydrate disassociation that destabilized the region, likely multiple times (e.g., Solheim et al., 2005). The slide itself though was most likely triggered by an earthquake (Bryna et al., 2005). Factors reducing slope stability may eventually lead to failure without other triggers, however, such failures are random, impossible to predict, and are unlikely to be regional. As all of the triggering mechanisms may trigger turbidity currents and are inherently difficult to distinguish, how can earthquake-triggered turbidites be distinguished from other turbidites? An equally important question is whether environments can be found that favor preservation of earthquake deposits, while disfavoring others? Essentially two methods can be
used to differentiate earthquake-generated turbidites from those originating from other processes: (1) Sedimentological examination; and (2) Tests for synchronous triggering of multiple turbidite systems that can eliminate nonearthquake origins. Both of these methods may be augmented by a historical earthquake record and land paleoseismic data if available.

In the following sections, we discuss these two methods and their global application, followed by specific applications to Cascadia, the Iberian margin, Japan, Sumatra, and other localities.

2B.4.1.1 Sedimentological and Mineralogical Characteristics

Japanese investigators have attempted to distinguish seismically generated turbidites (seismo-turbidites) from storm, tsunami, and other deposits. Nakajima and Kanai (2000) and Shiki et al. (1996, 2000a,b) argue that seismo-turbidites may in some cases be distinguished sedimentologically. Shiki et al. (2000b) carefully examined known seismo-turbidites in Lake Biwa, Japan, including the 1185 AD Lake Biwa/Kyoto earthquake (Mw = 7.4; Inouchi et al., 1996). These deposits are characterized by wide areal extent, multiple coarse-fraction pulses, variable mineralogical provenance (from multiple or line sources), greater organic content, greater depositional mass and coarser deposits than the barely visible storm-generated events (Figure 2B.16). They also concluded that defining the triggering mechanism of even known earthquake-related deposits was problematic, and that further study was needed. Nakajima and Kanai (2000) observe that a known seismo-turbidite from the 1983 Japan Sea earthquake caused multiple slump events in many tributaries of a canyon system, resulting in multiple coarse sediment pulses. The stacked multipulsed turbidite subunits had distinct mineralogies and were found deposited in order of travel time to their lithologic sources, demonstrating synchronous triggering of multiple parts of the canyon system (Nakajima and Kanai, 2000). Goldfinger et al. (2007) found a similar relationship with vertical stacking of separate mineralogic sources along the Northern San Andreas Fault. Gorsline et al. (2000) find that complexity, thickness, and areal extent also serve to distinguish Holocene seismo-turbidites in the Santa Monica and Alfonso Basins of the California borderland and Gulf of California, respectively. In the Santa Monica Basin, both flood generated and earthquake-generated turbidites are present. The flood turbidites are one-tenth to one-fifth the volume of the earthquake-generated events, which are more widespread. Similarly, turbidites in the Alfonso Basin were also found to be thicker and greater in aerial extent when earthquake generated. Gorsline et al. (2000) argued that reasonable estimates of discharge, sediment input, and source area can be used to constrain the sediment budget for flooding episodes to define upper bounds for what could be available for nonseismic turbidites.

2B.4.1.2 Distinguishing Hyperpycnal Underflows

Hyperpycnal flow is the density driven underflow from storm flood discharge of rivers into marine or lacustrine systems, and has been observed to generate turbidity currents, and proposed as a link to turbidity currents in a variety of settings. Hyperpycnites are commonly reported to have reverse-then-normal grading stemming from the waxing then waning nature of flood events (Figure 2B.17). The literature includes several reported cases and compares them to normally graded failure deposits such as those in the Var system (Mulder et al., 2001), Lake Biwa (Shiki et al., 2000a), and the Toyama deep sea fan (Nakajima, 2006). The dynamics of longitudinal and temporal variability and their effects have been discussed in detail by Kneller and McCaffrey (2003), and Mulder et al. (2003).

In some proximal settings such as large lakes, shelf basins, and fjords, records of both earthquakes and flood deposits have been found. In one of the best comparisons, St.-Onge et al. (2004) show that details of both seismic and hyperpycnal deposition in the Saguenay Fjord in eastern Canada are diagnostic, and argue that hyperpycnal deposits are distinguished by reverse grading at the base, followed by normal
grading. The diagnostic reverse-then-normal grading for hyperpycnal deposits has been widely reported and is attributed to waxing, then waning flow associated with the storm, although the waxing portion may later be eroded during later peak flows (Guyard et al., 2007). In the Saguenay Fjord, six events have normal grading alone and are inferred to be earthquake generated. Four others have similar basal units, but are topped by a reverse graded unit, and then a normally graded unit, with no evidence of hemipelagic sediment between the multiple units. These events are interpreted as an earthquake, followed by a hyperpycnite that resulted from the breaching of a landslide dam caused by the original earthquake. Dam breaching is a variant of the more common hyperpycnal scenario involving waxing and waning depletive flow (Kneller, 1995), but would likely result from a similar flow hydrograph (St.-Onge et al., 2004).

Documentation of hyperpycnal flows into lakes and shelf basins is abundant; however, evidence of such flows entering canyons systems and moving into deep water is relatively sparse. Most, if not all examples involve short distances between the river mouth and canyon head, either during Pleistocene low-stand conditions or in systems that have very narrow shelves during high-stand conditions. Hyperpycnal flows extend further from river mouths with high discharge (Alexander and Mulder, 2002), but documentation is sparse. Wright et al. (2001) observe that hyperpycnal flow is strongly affected by ambient currents and generally delivers sediment to the slope only upon relaxation of longshore currents. Most investigators

Figure 2B.16: (A) Soft X-radiograph (negative) showing sedimentary structures of upper 70 cm of core GH93-816 with a log of the core. Subdivisions T1–T5 represent turbidites while H1–H5 represent hemipelagites. Subdivisions (A) and (B) represent amalgamated beds within turbidites. Tb: parallel laminated sand; Tc: cross laminated sand/silt; Td: parallel laminated silt; E1: laminated mud; E2: graded mud. The lower part of the T5 bed and the B-Tm tephra layer has been disturbed by a coring effect. (B) Description of upper 70 cm of core GH93-816 showing sand content, water content, median grain size, sand composition, chemical composition and 137Cs concentration. Median grain size in phi. (I)–(III) in median grain size column represent sampling points for grain size distributions shown in Figure 2B.9. From Nakajima and Kanai (2000). Reprinted with permission from Elsevier.
cite Pleistocene examples or examples with little or no shelf width when referring to flows reaching the abyssal plain or lower fan reaches (e.g., Normark et al., 1998; Piper et al., 1999; Mulder et al., 2003; Normark and Reid, 2003). This is an expected result of sea-level change, or the near direct connection between a river and a canyon in the case of narrow shelves. Under low-stand conditions, rivers and canyons are more directly connected, and such flows are expected to dominate sediment delivery to the deep sea.

Thus the deep water deposition of hyperpycnites is closely coupled to sea-level control, or alternatively to climate shifts. An example of high-stand hyperpycnal flow has been reported for the Var River, in which the canyon and river mouth are less than 1 km apart (Mulder et al., 1998; Klaucke et al., 2000). Many large river systems deposit most of their load in river mouth bars, with lesser quantities making it past such bars in to a delta front slope (e.g., Yellow River, Li et al., 1998). Many canyon systems on continental margins were largely incised during Pleistocene sea-level low-stands (e.g., McNeill et al., 2000; Curray et al., 2002; Evans et al., 2005; LeRoux et al., 2005).

A good example is the 1969 El Nino flood, which input ~25 million tons of sediment (5× the present yearly Columbia River sediment load; Sherwood et al., 1990) to the Santa Ana River in southern California over a 24 h period, in close proximity to nearby canyon heads (Drake et al., 1972). Sediment from this extreme flood did not continue down canyons as hyperpycnal flow, but deposited as a distinct yellow unit on the shelf and upper slope. Over the next 10 years, the flood sediment moved

**Figure 2B.17:** (A) Idealized stratigraphy resulting from hyperpycnal flow, characterized by a coarsening upward sequence followed by a fining upward sequence attributed to a waxing then waning hydrographic profile during a storm event. Other events with a similar hydrographs, such as a gradual dam breaching may produce similar stratigraphy (after Mulder, 2001). (B) Typical stratigraphic sequence from a turbidite with multiple fining upward pulses from core M9907-12PC in Juan de Fuca Channel, Cascadia margin. This turbidite and nearly all others in the Holocene Cascadia Basin turbidite sequence exhibit multipulsed stratigraphy, with no waxing phase. Multiple fining upward sequences are capped by a fine mud tail signaling the final waning of the turbidity current.
downslope as turbid layer transport caused by storm wave resuspension, and deposited as yellow layers between varves of the Santa Barbara Basin (Drake et al., 1972).

Hyperpycnites are also commonly organic rich as compared to seismic turbidites, having their sources in floods rather than in resuspension of older canyon wall material as in earthquake triggering (Shiki et al., 1996, 2000b; Nakajima and Kanai, 2000; Mulder et al., 2001). It has been suggested that this distinction may be used as a basis for distinguishing earthquake and storm deposits using OSL dating (Shirai et al., 2004). However, we suspect that this generalization may easily be violated as in the case of floods in very arid regions, or earthquakes in heavily vegetated areas. For example, because the river drainage basins feeding Cascadia Basin are heavily vegetated, the Holocene turbidite tails linked to earthquake origins through a variety of methods (Goldfinger et al., 2008, 2009) are characterized by significant quantities of plant fragments (Nelson, 1976).

Whether hyperpycnal flows can reach deep water via canyon systems incised during the sea-level low stands appears to be a function of shelf width, steepness, river peak storm discharge, high-stand aggradation, and the wave and current climate during peak storm discharge. However, the requirements for and evidence of hyperpycnal flows to the deep ocean under high-stand conditions (excepting very narrow shelves) remain poorly known at best (Mulder et al., 2001). A well-documented example for the Toyama Channel and fan is given by Nakajima (2006) in which long-traveled and long-lived pulsed flows traveled 700 km to a deep sea fan. As with other examples, no shelf width buffers the river source from the canyon channel system in the Toyama system. In cases of narrow shelves, a turbidite record in an offshore basin or abyssal plain may well contain a mixture of hyperpycnal, sediment failure, and earthquake-generated turbidites. For systems that minimize these effects, those with wide continental shelves, or topographic barriers isolating the slope and abyssal plain the turbidite record is more likely to contain a dominantly earthquake record (Nakajima and Kanai, 2000; Abdeldayem et al., 2004; Goldfinger et al., 2008, 2009). The implication is that caution must be exercised to examine the river systems, their relationship to sea level during periods of interest, and the physiographic conditions of shelf width, forearc basins, and other barriers to hyperpycnal flow when evaluating a particular setting for turbidite paleoseismology.

### 2B.4.1.3 Synchronous Triggering

While there are few definitive sedimentological studies linking earthquakes directly with turbidites on the basis of the deposits themselves, most studies have focused on aspects of earthquake processes that are unique, and therefore eliminate most or all of the turbidite triggering mechanisms other than earthquakes. The primary characteristic that can easily be distinguished in sediment cores is spatial extent. When turbidite deposits can be correlated among widely spaced sites, synchronous deposition can be established or inferred, and if the spatial extent exceeds that reasonable for other mechanisms, then earthquake triggering is likely. Virtually all studies that make the linkage between earthquake triggering and turbidites invoke this test in some fashion, including those cited previously under sedimentological examination (Adams, 1990; Nakajima and Kanai, 2000; Gorsline, 2000; Goldfinger et al., 2003b, 2007, 2008).

### 2B.4.1.4 Numerical Coincidence and Relative Dating Tests

In his synthesis of Cascadia Basin turbidite events, Adams (1990) observed that in several canyons feeding into a confluence, cores contained 13–14 turbidites above a regional tephra, the Mazama ash. Below the confluence, cores in the main Cascadia channel also contained 13 turbidites (Figure 2B.18). He reasoned that these events must have been synchronously triggered because if these events had been independently triggered with more than a few hours separation in time, cores taken below the confluence
should contain from 26 to 28 turbidites, not 13 as observed. The only alternative is that 13 turbidites also dropped out of the sequence due to the more distal position of the downstream core, an unlikely coincidence. The importance of this simple observation is that it demonstrates synchronous triggering of turbidity currents in tributaries the headwaters of which are separated by 50–150 km. The synchronicity demonstrated by this “confluence test” is also supported by the similar numbers of events alone, without the existence of the confluence, suggesting either synchronous triggering, or a regionally coherent coincidence. Off the California margin, Goldfinger et al. (2007) demonstrate that turbidites adjacent to the Northern San Andreas Fault also converge at a number of channel confluences and follow a similar pattern to that observed in Cascadia, remaining constant in number above and below the confluence.
Chapter 2B Sub-Aqueous Paleoseismology

2B.4.1.5 Stratigraphic Correlation

The lithostratigraphic correlation of turbidite stratigraphy offers a straightforward method to test for event synchrony. The detailed geophysical “fingerprinting” of turbidites through their grain size distributions and other physical properties has direct implications for synchronous origins of the deposits. Geophysical signatures, commonly in the form of density, magnetic susceptibility, velocity, XRF composition, and other parameters can serve to establish a stratigraphic fingerprint (Figure 2B.19). Goldfinger et al. (2007, 2008, 2009) found that these “fingerprints” can be persistent among sites and over considerable distances. These fingerprints sometimes retain a remarkable similarity at sites along strike, but also commonly evolve somewhat along strike and downchannel in subtle ways that can be traced from one site to another. That such grain-size “fingerprints” exist suggest that triggering mechanisms that produced them, or the hydrodynamics of the separate canyon systems must have some commonality, as producing matching grain size patterns by coincidence is unlikely. Goldfinger et al. (2008, 2009) observe that in Cascadia the individual stratigraphic signatures can be traced across multiple canyon/channel systems, and at least one slope basin. Some of these sites have no physical connection, and the basin site is isolated from all other sites and sources of fluvial input.

2B.4.2 Turbidite Paleoseismology

2B.4.2.1 Cascadia

Goldfinger et al. (2003, 2008, 2009) investigated turbidite systems located on the continental margin of Cascadia Basin from Vancouver Island, Canada to Cape Mendocino California, USA. Cascadia Basin contains a variety of types and scales of turbidite systems including multiple canyon sources on the Washington margin that funnel turbidites into Cascadia Channel (1000 km length); Astoria Canyon on the northern Oregon margin that feeds Astoria submarine fan (300 km diameter) containing channel splays with depositional lobes; Rogue Canyon on the southern Oregon margin that feeds a small (<5 km) base-of-slope apron, and Trinidad, Eel, and Mendocino canyons (30–100 km length) on the northern California margin that feed into plunge pools, sediment wave fields, and channels. Detailed swath bathymetric data and core sampling procedures verify that key turbidite channel pathways of Cascadia Basin are open and provide a good turbidite event record. Proximal canyon mouth and inner fan channel areas have erratic turbidite event records because of extensive cut and fill episodes in turbidity currents; however, even in these difficult locations, complete records can be found in some point bars, terraces and canyon walls that are slightly elevated above the channel thalweg. The most consistent turbidite event records occur in distal locations of continuous deep-sea channel systems such as Cascadia Channel.

Multiple tributary channels with 50–150 km spacing and a wide variety of turbidite systems with different sedimentary sources contain 13 post-Mazama ash and 19 Holocene turbidites in Cascadia Channel, Juan de Fuca Channel off Washington, Hydrate Ridge slope basin, and Astoria Fan off northern and central Oregon. All of these events are also recorded on Rogue Apron of southern Oregon, with the addition of smaller local events recorded as silt or mud turbidites. Nineteen Holocene turbidites are found along the northern and central margin and are recorded in southern cores with 22 interspersed smaller events.

Goldfinger et al. (2008, 2009) used 14C ages, the previously described “confluence test,” and stratigraphic correlation of turbidites to determine whether turbidites deposited in separate channel systems were correlative and pass tests of synchronous deposition to test for earthquake origin (Figure 2B.20). The confluence test shows that a coherent record of 19 Holocene turbidites pass this test along the northern
Figure 2B.19: Correlation details from two representative pairs of cores on the Cascadia margin. (A) Events 8–11 in cores from Juan de Fuca Channel (left) and Cascadia Channel (right). Left traces are raw gamma density, right traces are magnetic susceptibility. Lithologic logs are also shown. Note correspondence of size, spacing, number of peaks, and trends of physical property traces between these cores. (B) Similarly displays events T10–T14 in Juan de Fuca Channel (left) and T10d–T14 in Rogue Channel (right). (A) Cores are part of the same channel system, distance along channel = 475 km. (B) Cores are in channels that do not meet, separation distance = 500 km. Note that correlation of longer sections and $^{14}$C data show that T10f and T10 do not correlate in (B). Similarly, Mazama ash appears in T14, not T13 in Rogue apron, see text for discussion.

Modified after Goldfinger et al. (2008). (See Color Insert.)
margin. This record represents the entire record of Holocene turbidites along the northern Cascadia margin, leaving no deposits from other sources.

Another key piece of evidence to address multiple triggering mechanisms is the correlative turbidite sequence from Hydrate Ridge Basin at ~44.5N on the Oregon lower slope. This slope basin is completely isolated from land sources of sedimentation, being surrounded by ridges 500–1800 m above the basin floor that prevent downslope transport into the basin from any source other than the flanks of the ridge itself. The physiography and great depth of the basin eliminate input from storms, tsunami, hyperpycnal flow and other external sources, as evidenced by the absence of Mazama ash. The turbidite record from this key site is correlated to other margin sites on the basis of stratigraphic “fingerprints” and 14C ages. The strong correlations to this site comprise an independent test of turbidite triggering at a site where all triggers save earthquakes and self failure are eliminated by the local physiography.

The synchronicity of a 10,000 year turbidite event record for 500 km along the northern half of the Cascadia Subduction Zone is best explained by paleoseismic triggering by great earthquakes. The southern Cascadia margin includes correlated additional events, many of which are also correlated to Hydrate Ridge Basin, though there are no channel confluences that can be used to test for synchronicity. The average Holocene great earthquake recurrence was found to be ~500 years, for the northern margin, similar to the onshore rate. Goldfinger et al. (2009) report that the recurrence times and averages are also supported by the thickness of hemipelagic sediment deposited between turbidite beds. Using stratigraphic correlation and 14C ages, they report that the southern Cascadia margin can be divided into at least three seismic segments that include all of the northern ruptures, as well as ~22 thinner turbidites of restricted latitude range that are correlated between multiple sites. At least two northern California sites, Trinidad and Eel Canyons, probably also record numerous small sedimentologically or storm-triggered turbidites, particularly during the early Holocene when a close connection existed between these canyons and associated river systems under lowered sea-level conditions.

The combined stratigraphic correlations, hemipelagic analysis, and 14C framework suggest that the Cascadia margin effectively has four rupture modes: 19 full or nearly full-length ruptures; two or three ruptures comprising the southern 50–70% of the margin, 9 or 10 events including the southern 50% of the margin and 9 events restricted to southern must Oregon and northern California (Figure 2B.20). The shorter rupture extents and thinner turbidites of the southern margin correspond reasonably well with spatial extents interpreted from the onshore paleoseismic record, supporting margin segmentation of southern Cascadia. The total of 41 events defines a Holocene recurrence interval for the southern Cascadia margin of ~240 years.

Goldfinger et al. (2009) report that turbidite physical properties along the Cascadia margin reveal a consistent record of turbidite mass per event along the northern margin for many events. Larger turbidites also have a moderately good correlation with the time interval following each event and are uncorrelated with the preceding time. They infer that larger turbidites likely represent larger earthquakes, and therefore the correlation with following time intervals suggests that Cascadia full margin ruptures may follow a time-predictable earthquake model. The long paleoseismic record also apparently indicates a repeating pattern of clustered earthquakes that includes three Holocene cycles of five earthquakes followed by an unusually long interval of 700-1000 years.

Goldfinger et al. (2009) suggest that the pattern of long time intervals and longer rupture for the northern and central margin may be a function of high sediment supply on the incoming plate smoothing asperities and potential rupture barriers. The smaller southern Cascadia segments correspond to reduced sediment supply and potentially greater interaction between lower plate and upper plate heterogeneities.
C. Goldfinger

[Diagram and explanation text]

Explaination:
- Core break
- Magnetic hi-res for PCs and KC, loop mag for TCs
- Density (g/m^3)
- Pb 210 activity
- Stratigraphic correlation, major
- Stratigraphic correlation, minor

2600 Calibrated 

2600 Hemipelagic age calculated from event before at one or both; BP 1990
2600 Suspect or no known benthic age; BP 1990
Goldfinger et al. (2009) make comparisons of the frequency of other potential triggering events, including bolide impacts, distal tsunami, storms, and crustal earthquakes and infer that these mechanisms are unlikely to be responsible for the observed record. During great earthquakes, on the other hand, the entire canyon system is affected, a canyon length that can exceed 100 km in Cascadia. The rupture zone also underlies the full length of all of the Cascadia canyons at a shallow depth, creating a nearly ideal setting for triggering slope failures. During a great earthquake, the hypocentral distance to locked fault is never more than between 2 and 10 km from the canyon walls, which likely fail in nearly continuous wall failure during the severe ground shaking of a large earthquake. Peak ground accelerations at such short distances to a great subduction earthquake can be estimated using the attenuation relationships of Atkinson and Boore (1997) and Youngs et al. (1997) to between 2 g (Youngs et al., 1997) soil sites and 3.5 g (Atkinson and Boore, 1997) for rock sites. This represents a tremendous suspension and liquefaction force far greater than anything possible from surface ocean waves.

Finally, the recurrence intervals of Cascadia Basin offshore turbidites (Trinidad, Eel, and Mendocino channels excepted) closely match that of the onshore paleoseismic record (Goldfinger et al., 2003a,b, 2007, 2008, 2009). The lack of turbidites overlying the most recent turbidite, dated to within a decade of the 1700 AD Cascadia earthquake indicates that no other triggering mechanism has produced an observable turbidite in the last 300 years, except in Trinidad and Eel Channels with narrow shelves and a local river source. The lack of turbidite triggering in Cascadia Basin by historic El Nino storm and flood events (1964, 1998–1999), and the 1964 Alaskan earthquake tsunami suggest that storm events and tsunami, whether or not sediment is transported to canyon heads, do not generally result in correlative abyssal plain turbidites. The mean peak AMS age of 230 (140–340) cal BP from four channel systems for the youngest turbidite event in Cascadia Channel T1 differs by only 15–20 years from (1) the coastal

Figure 2B.20: (A) Holocene rupture lengths of Cascadia great earthquakes from marine and onshore paleoseismology. Four panels showing rupture modes inferred from turbidite correlation, supported by onshore radiocarbon data. (A) Full or nearly full rupture, represented at most sites by 20 turbidites, though with greater uncertainty in southern extent (we include Pleistocene T19 in the figure, but not in the statistics). (B) Mid-Southern rupture, represented by two (1?) events. (C) Southern rupture from central Oregon southward represented by 9 (10?) events. (D) Southern Oregon/northern California events, represented by eight events. Southern rupture limits vary with each event, and many events older than ~5000 years are limited by lack of core older data. Dashed white line offshore indicates reduced confidence in correlations south of Trinidad Canyon. Recurrence intervals for each segment shown in left panel. Each segment includes all full margin events, plus those exclusive to that segment. Rupture terminations are approximately located at three forearc structural uplifts, Nehalem Bank (NB), Heceta Bank (HB), and Coquille Bank (CB). Paleoseismic segmentation shown is also compatible with latitudinal boundaries of Episodic Tremor and Slip (ETS) events proposed for the downdip subduction interface (Brudzinski et al., 2007). These boundaries are shown by white-dashed lines. A northern segment proposed from ETS data at ~48N does not appear to have a paleoseismic equivalent. (B) Correlation plot of Holocene marine turbidite records and 14C ages along the Cascadia margin from Barkley Channel to Eel Channel. All cores are vertically scaled to match Rogue core 31PC which is at true scale. Turbidite ages are shown using probability peaks and averaged where multiple ages at one site are available. Turbidites linked by stratigraphic correlations are shown by connecting lines. Full margin events correlated by using stratigraphy and 14C are shown thicker, local southern Cascadia events are thinner and dashed. Modified after Goldfinger et al. (2009). (See Color Insert.)
paleoseismic ages that consistently center about 250 cal BP (AD 1700; Nelson et al., 1995) and (2) the
Japanese tsunami evidence showing a date of January 26, 1700 for the youngest great earthquake on the
Cascadia Subduction Zone (Satake et al., 1996, 2003). This further validates the synchronous turbidite
event record and associated high-resolution AMS radiocarbon ages as a method to provide a long-term
paleoseismic record. Temporal correspondence between the onshore and offshore paleoseismic records
along the Cascadia margin is quite good, despite a variety of methods and lines of evidence onshore.
Within the time ranges that the two records overlap, there are few significant discrepancies (Goldfinger
et al., 2009). The ties between onshore and offshore paleoseismic data remain limited to radiocarbon
timing for all sites except Effingham Inlet on Vancouver Island, which contains turbidites with possible
stratigraphic correlatives offshore.

Goldfinger et al. (2009) conclude that turbidite systems of the Cascadia Basin are an ideal place to
develop a turbidite paleoseismologic method and record because: (A) a single subduction zone fault
underlies the Cascadia submarine canyon systems, (B) multiple tributary canyons and a variety
of turbidite systems and sedimentary sources exist to use in tests of synchronous turbidite triggering;
(C) the Cascadia trench is completely sediment filled, allowing channel systems to trend seaward across
the abyssal plain rather than merging in the trench, (D) the continental shelf is wide, favoring disconnection
of Holocene river systems from their largely Pleistocene canyons, and (E) excellent stratigraphic datums,
including the Mazama ash (MA) and a distinguishable Holocene/Pleistocene boundary (H/P), are present
for correlation of events and anchoring the temporal framework in turbidite systems within the northern
two thirds of the basin.

2B.4.2.2 Marmara Sea

Correlating turbidites with the historical record is a good way to begin testing the turbidite record for
seismic origin if a historical record is available. Considerable effort has been directed toward the Marmara
Sea following the 1999 Izmit earthquake to map the submarine. North Anatolian Fault (NAF), concern has
been heightened because the fault segment immediately to the west of the 1999 rupture may fail next, and
the close proximity of this segment to Istanbul represents a significant hazard to the city. McHugh et al.
(2006) describe work in the submarine pull-apart basins of the NAF within the Marmara Sea in which
CHIRP seismic profiles, multibeam bathymetry and cores were used to test the connections between the
turbidite basin fill and the NAF. Unlike channel settings used in Cascadia and the NSAF, the NAF work
used cores sited in local depocenters along the fault. The sedimentation rates in the Marmara basins are
quite high (0.5–1.0 cm/yr) making possible the resolution of events spaced closely in time. The turbidites
in the deep Marmara basins, close to the fault segments are differentiated to some degree from thinner
bedded turbidites on the shelf and slope, which McHugh et al. attribute to climatic events such as floods,
though the earthquake and climatic records are most likely mixed. Thick Holocene deposits (5–20 m)
were found on the basin flanks and presumably fail into the basin during earthquakes. In this setting,
the method of emplacement and turbidite pathways are not completely clear, though the sedimentary
packages thicken basinward, and the fining upward sequences require transport into the depocenters from
upslope. Much like the Cascadia and NSAF turbidites, the cores revealed turbidites with multiple fining
upward sequences capped by fining upward silt and hemipelagic foram-rich clay (the authors chose to use
the term “homogenite” for these deposits which closely fit the description of turbidites, though they have
long mud tails, see Shanmugam, 2006). These deposits were dated with a combination of $^{137}$Cs and
radiocarbon to search for matches for the historical series of nine earthquakes in 181 AD, 740 AD, 1063
AD, 1343 AD, 1509 AD, 1766 AD, 1894 AD, 1912 AD, and 1965 AD (Figure 2B.21). In the basin
depocenters, the thinner bedded turbidites of the shelf and upper slope were not present. In cases such as
Cascadia where the historical record is limited or nonexistent, the burden of demonstrating earthquake
origin is a relatively complex series of tests and regional correlations as previously described. In the case where a good historical record is available, however, the establishment of a good ground truth for seismic turbidites is much simpler. The cores McHugh et al. (2006) report in the Ganos, Tekirdağ, and Central Basins appear to record most, but not all of the local earthquakes on the NAF. The core record also includes some of the events in the Sauros basin, some 60 km distant, bounded by a different fault segment. Cores collected from shallow water near the NAF west of the Hersek peninsula also contained disturbance deposits that were tentatively related to three historical earthquakes in 1509, 1766, and 1860.

Figure 2B.21: Turbidite paleoseismologic example from the Marmara Sea. Grain size variability ranging from fine-sand to fine-silt and an increase in the total organic carbon (TOC %) of the sediments permit resolution of three homogenite (turbidite) deposits (40–44, 60–92, 100–112 cm). The homogenites are initiated by a sharp basal contact overlaid by multiple sand and silt-size laminae that fine upward to a thick wedge of homogenous fine-grained silt. The homogenites are separated by thin beds of clay (5–10 cm). Short-lived radioisotopes and radiocarbon chronology permit constructing an age model for correlation of the homogenites to the historical record of earthquakes: the large 1912 Ganos event $M_s > 7$ that lead to the deposition of a 30 cm thick homogenite and two smaller events that occurred in the Gulf of Saros. Sedimentation rates of 1 cm/yr were calculated for the upper 40 cm of the core that is apparently undisturbed. From McHugh et al. (2006). Reprinted by permission of Elsevier. (See Color Insert.)
Additional work described in Beck et al. (2007) describes work in progress on two giant piston cores collected in the Marmara Central Basin and reaches a somewhat more complex conclusion about the sedimentary section and the role of earthquake-generated turbidites from the late deglacial through the Holocene. The long (26 and 37 m) cores collected with this system appear to include the entire marine history of the central basin, as well as a previous lacustrine episode. Beck et al. (2007) suggest that while much of the basin fill is likely earthquake generated, they observe several features for which the explanation is perhaps not so straightforward. One of the turbidites is extremely large, and properly called a megaturbidite with an upper homogenite unit /C24 m thick. This unit roughly corresponds with the transition from lacustrine to marine conditions. They also observe a significant decrease in turbidite frequency at about this same time, ~16,000 yr BP. Goldfinger et al. (2007) also observe occasional turbidites among regional correlatives that are likely earthquake generated that are inexplicably outsized by comparison to others, and to turbidites generated by the maximum regional earthquake, the 1906 San Andreas event. Interestingly, Beck et al. (2007) propose that finely laminated units may be the result of bottom seiche currents linked to significant earthquakes. Finally, Beck et al. (2007) suggest that observations of planar fluid escape structures may represent liquefaction from ground shaking, as they appear to be spatially correlated to the turbidites in the cores, though they could potentially be related to coring with the large Calypso system.

2B.4.2.3 Northern San Andreas Fault

Using similar methods to their Cascadia work, Goldfinger et al. (2007) used 74 piston, gravity and jumbo Kasten cores from channel and canyon systems draining the northern California continental margin to investigate the record Holocene turbidites along the adjacent Northern San Andreas Fault. This fault is offshore or near the coast from San Francisco to the Mendocino Triple Junction and apparently close enough to offshore canyon heads to trigger turbidity currents. The late Holocene turbidite record off northern California was found to pass tests for synchronous triggering and was correlated using multiple proxies between numerous sites from Noyo Channel near the triple junction and the latitude of San Francisco. Preliminary comparisons of the temporal event record based in 14C ages with existing and in progress work at onshore paleoseismic sites show good correlation, further circumstantial evidence that the offshore record is primarily earthquake generated. During the last ~2800 years, 15 turbidites are recognized, including the one likely generated by the great 1906 earthquake. Their chronology establishes an average repeat time of ~200 years, similar to the onshore value of ~230 years. Along-strike correlation suggests that at least eight of the youngest 10 of these events likely ruptured the 320 km distance from the Mendocino Triple Junction to near San Francisco.

The long paleoseismic histories developed for the adjacent Cascadia and NSAF systems allowed Goldfinger et al. (2008) to relate the NSAF paleoseismic history to the similar dataset from the Cascadia (Figure 2B.22). They note that the recurrence interval for the NSAF is quite similar to that of southern Cascadia, where the combined land and marine paleoseismic record includes a similar number of events during the same period. While the average recurrence interval for full margin Cascadia events is ~500 years, the southern Cascadia margin has a repeat time of ~220 years during the most recent 3000 year period, similar to that of the NSAF. Comparing these two records in several ways, using offshore data, land data, and the combined land–marine average, they find that 12 of the 15 NSAF events apparently occurred in close temporal proximity to Cascadia earthquakes. There appeared to be a slim temporal lag of ~0–80 years, averaging 25–45 years, with Cascadia preceding the NSAF, (as compared to ~80–400 years by which Cascadia events follow the NSAF).
Figure 2B.22: OxCal age model for the youngest 15 events in the NSAF offshore system, and comparison to onshore NSAF ages. Cascadia OxCal PDFs are shown in blue, with lighter blue used where only Hemipelagic ages are available. Land ages from OxCal combines are shown in red. Cascadia mean event ages are also shown with blue arrows for well-dated turbidite events, Purple arrows for hemipelagic age estimates, and light red arrows for onshore paleoseismic events. See text for discussion and tables for data used and criteria, and discussion of temporal relationships. Inter-event times based on hemipelagic sediment thickness (represented by gray segments of NSAF PDFs) were used to constrain original 14C calendar age distributions (gray traces) using the SEQUENCE option in OxCal. Inter-event times were estimated by converting hemipelagic sediment thickness...
Based on the observed temporal association, Goldfinger et al. (2008) modeled the coseismic and cumulative postseismic deformation from great Cascadia megathrust events and compute-related stress changes along the NSAF to test the possibility that Cascadia earthquakes triggered the penultimate, and perhaps other NSAF events. They concluded that the Coulomb failure stress (CFS) resulting from viscous deformation related to a Cascadia earthquake over ~60 years does not contribute significantly to the total CFS on the NSAF. However, the coseismic deformation increases CFS on the NSAF by up to about nine bars following a typical Cascadia earthquake, most likely enough to trigger that fault to fail in north-to-south propagating ruptures (Figure 2B.22).

2B.4.2.4 Kurile Trench

Marine turbidites as paleoseismic recorders have been investigated along the Japanese islands, primarily in the trench systems and submarine canyons along the eastern coast. Along the eastern Hokkaido forearc along the Kuril Trench, Noda et al. (2004, 2008) have investigated the turbidite stratigraphy in Kushiro submarine canyon, offshore Kushiro. The earthquake history in this region during the last few centuries is well known from the historical literature. Two gravity cores (GH03-1033 and GH03-1034) were collected from the bottom of the canyon, and contain a number of turbidites. Sedimentological, geochemical, and micropaleontological data as well as high-resolution seismic data have been used to identify character, provenance, and recurrence intervals of the canyon turbidites. Three tephras from known volcanic events during AD1739, AD1694, AD1667, and AD1663 were used to develop the age model of the cores and recurrence intervals of the turbidites. In the upper canyon, thick mud in the channel suggests that the upper canyon has not been an active pathway during the Holocene. The middle canyon core (GH03-1034) had a source material inferred to be the upper canyon walls on the basis of sand composition and benthic foraminiferal analysis. A recurrence interval of 68 years for the late Holocene is similar to the historical rate of 79.7 years. Individual turbidites were also found to correspond well to the known historical earthquake record (Noda et al., 2004). There are several seismic segments within the Kurile Trench. Noda et al. (2008) reported a recurrence interval less than 113 years for another segment along the Kurile Trench.

2B.4.2.5 Nankai Trough

Along the eastern Nankai Trough, Ikehara and Ashi (2005) have observed turbidite sands in cores collected from slope and forearc basins in the Tokai region. Two cores contain 15 and 13 turbidites from two slope basins along the Tokai Thrust, an out-of-sequence thrust in the Nankai accretionary prism. The turbidite ages, determined by 14C dating of planktonic foraminifera suggest that during the last 3000 years, turbidite frequency along the eastern Nankai Trough is 100–150 years, similar to the known
intervals of large interplate earthquakes from the historical and archeological records. Ikehara and Ashi (2005) also report occurrence of turbidites in the northern Kumano Trough with a recurrence frequency of \(~200–250\) years, or about twice as long as the known interval for interplate earthquakes along the eastern Nankai Trough. Turbidites in a core from the Omine Ridge, an outer ridge near the Kumano Trough near an out-of-sequence thrust, suggest a \(~1000\) year recurrence. The inconsistent results from Nankai point out the importance of spatial coverage to test for earthquake origin among turbidite records, and the variable results that may be found in different settings within the Nankai and other subduction zones (Ikehara and Ashi, 2005).

2B.4.2.6 Sumatra

The December 2004 Sumatra–India earthquake and tsunami represents an opportunity to catalogue marine effects from a very well-recorded series of events, many of which are unknown or poorly known at present. A suite of piston, gravity, Kasten, and multicores was collected along the length of the Sumatra margin, from the 2004 rupture zone in the north, to the southern tip of Sumatra Island (Patton et al., 2007). Preliminary work suggests that like Cascadia, stratigraphic correlation may be possible in the Sumatra area, supported by numerous tephras with distinct compositional signatures.

Because there were no opportunities for a “confluence test” along the Sumatran margin, their strategy was to densely sample both trench and basin sites to test correlations between these two disparate and isolated site types to test for earthquake origin. Preliminary analysis suggests that the cores contain turbidites most likely generated by the 2004 and 2005 northern Sumatra great earthquakes. These are represented by a large shallow multipulse event overlain by a smaller single pulse event at the seafloor, with no observed hemipelagic sediment between them. Patton et al. (2007) suggest that other turbidites correlate over distinct strike lengths, indicating that seismic segmentation may be resolvable with this dataset. Ongoing \(^{14}\)C and \(^{210}\)Pb dating with stratigraphic correlation will test the origins and connectivity of these and numerous other Holocene turbidites along this poorly known subduction margin.

2B.4.3 Offshore Tsunami Deposits

Evidence of onshore tsunami deposits exists in many forms, many of which clearly have utility as paleotsunami and paleoearthquake records, while others are poorly known and somewhat speculative. Potential modes of tsunami deposition in the marine environment include tsunami-related sedimentation in bays, lagoons, and lakes whose seaward boundaries were overwashed by tsunami waves (lacustrine deposits are discussed more fully in Section 2B.4.4). (We focus here on earthquake-generated deposits at the expense of significant work that has been done in impact generated tsunami, particularly hotly debated work near the KT boundary, e.g., Smit et al., 1996; Keller et al., 2003).

Lesser known than onshore deposits are offshore deposits in open bays, shelves, and forearc basins that result from tsunami passage and backwash. A very few well-documented cases have been reported, and these types of deposits remain to some extent in the realm of speculation. Shiki et al. (2008) and Shiki and Tachibana (2008) discuss the conceptual issues and problems surrounding tsunamites as well as their importance to the geologic record, and relationship to climate and tectonic cycles and events in some detail. Shiki et al. (2008) review the features expected in submarine tsunamites and the sedimentary structures related to various parts of the tsunami wave train. Shiki and Yamazaki et al. (1996) discuss a potential tsunamiite in the upper bathyal Miocene section onshore in central Japan that includes cobble imbrication from high-flow velocities and an association with probable shaking evidence, a key discriminator for tsunamiites which are otherwise difficult to distinguish (e.g., Dawson, 1999).
A somewhat similar deposit assigned to shallow water/shoreface depths is reported in the Miocene of Chile (Cantalamessa and Di Celma, 2005). Numerous deposits are reported in the ancient onshore geologic record of other potential tsunamiites, though evidence for tsunami origin is generally not strong. Fujiwara et al. (2000) describe tsunami deposits in a drowned valley on the Boso Peninsula of Japan. He proposes a tsunami depositional model based on depositional structures, high-resolution grain size analyses and the taphonomy of molluscan shells and suggests that details of the tsunami waveform may be deduced from the stack of depositional units. Weiss and Bahlberg (2006) performed an analysis of storm and tsunami wave energy and preservation potential along the Australian coast. They used a combination of hydrodynamic modeling and a simplified Hjulstrom–Sundborg diagram and concluded that the most powerful storm and tsunami waves both produce conditions near and at the sea bed that allow the transport of similar sediment grain sizes, up to meters in diameter. The implication is that offshore tsunami deposits in that locality would be reworked by storm waves. For their site-specific study at Brisbane, they concluded that preservation of tsunami deposits is most likely at depths greater than 65 m. Larger tsunami such as those produced by impacts, very large submarine landslides (Goldfinger et al., 2000; McMurtry et al., 2004) may well overcome this problem, as might a selection of localities not subjected to significant storm wave influence.

The few reports of tsunami backwash deposits in nearshore environments are suggestive of potential for preservation, perhaps more subject to special conditions, but possibly offering sites where other modes of preservation are not available. van den Bergh et al. (2003) describe a shallow water tsunami deposit from the 1883 Krakatau eruption using textural, compositional and 210Pb geochronological data. The deposit is associated with the 1883 eruption tephra, and thus its origin is relatively clear. The deposit consists of a sandy layer with abundant reworked shell fragments and material apparently locally derived eroded from the seabed. Also note that the deposit included land-derived components when near the coast. Nearshore deposits (<50 m water depth) may also preserve critical information as to wave direction and speed, though these have not been reported to our knowledge. Such evidence might include the preservation of sediment aprons, sand bars, large sediment waves and debris layers deposited during backwash. Also, large objects (boulders, coral blocks, human artifacts) may be dragged or deposited on the seafloor, producing a debris field and other scattered evidence on the seafloor.

An unusual example of a potential hybrid deposit, part tsunami deposit, part turbidite has been described from the Mediterranean seabed. A widespread unit known as a homogenite has been widely described in the Ionian and Sirte abyssal plains and other scattered locations in the central Mediterranean. This deposit, up to several meters thick, is mostly homogenous clay to silt with little or no grading (Cita et al., 1984; Kopf et al., 1998). Cita and Aloisi (2000) describe a pelagic (Type A) and a shallow water (Type B) homogenite. The pelagic deposit, without indications of a shallow water source, has a coarse fraction (sand size) consisting only of planktonic foraminifers (Sironi and Rimoldi, 2005). This deposit has been attributed to a catastrophic eruption of Santorini volcano, which Sironi and Rimoldi (2005) suggest, generated a tsunami during the collapse of the Santorini Caldera. The tsunami in turn is thought to have destabilized mostly hemipelagic marine sediments on shallower ridges such as the Mediterranean and Calabrian Ridges, generating turbidity currents that deposited the homogenite on the abyssal plain. Sironi and Rimoldi (2005) further suggest that this homogenite was then overlain with a megaturbidite (up to 24 m thick) originating on the African continent as a result of the tsunami arrival. Pareschi et al. (2006) argue for the same origin for the homogenite, but argue for an origin from Mt. Etna rather than Santorini, and provide tsunami modeling to support the distribution of liquefaction potential to support their model.

While definitive assignment of deep ocean turbidites to a tsunami origin is rare, these papers suggest that such deposits are likely to exist in the geologic record. Along the Cascadia margin, Goldfinger et al.
(2009) calculate the potential for tsunami triggering of turbidites on the upper continental slope, and conclude that the potential exists, though correlation with onshore earthquakes suggests that the Cascadia Holocene turbidites are of local earthquake origin.

2B.4.4 Lacustrine Environments

2B.4.4.1 Lacustrine Sediment Pulses Caused by Earthquake-Generated Landslides

Adams (1980) measured sediment loads of rivers in New Zealand immediately following earthquakes and observed an order-of-magnitude increase in sediment load for a period of several months. He correlated increases in load in different areas with the density of earthquake-triggered landslides in those areas and concluded that seismically induced landslides generate large increases in fluvial sediment load, which, in turn, cause increases in sedimentation rates in lakes and oceans. These observations have been corroborated with published observations from earthquakes elsewhere (Adams, 1981; Dadson et al., 2004).

On the premise of these observations, Doig (1986) analyzed organic-free silt layers 0.3–2.0 cm thick in otherwise organic-rich lake sediment in eastern Canada. Using sedimentation rates and radiometric methods, three of these layers were correlated with known earthquakes of AD 1663, 1791, and 1860 ± 1870 (two events combined). Two older silt layers were likewise dated and attributed to paleoearthquakes in AD 1060 and 600. Doig (1986) stated that cores from deep lakes likely will yield the best cores for this type of analysis because of lack of bioturbation. He also warned that dating young (a few hundred years) silt layers characterized by lack of organic material can be difficult; he suggested that $^{210}$Pb and $^{137}$Cs are the ideal radiometric methods for this type of analysis (see details of dating techniques in Chapters 1 and 2A).

2B.4.4.2 Landslide, Turbidite, and Tsunami Deposits in Lakes

A number of lake deposits in various settings have been interpreted as related to seismic shaking, landslides into lakes, submarine landslides, and tsunami overwash into lakes. Lake settings may not offer the constant sediment supply of the offshore environment, but have the advantage of seasonal changes that may be reflected in annual sediment patterns, offering precise chronologies.

Alpine Lakes Beck et al. (1996) report evidence of liquefaction and differential compaction with rapid fluid escape (water and/or gas) in the form of ball-and-pillow structures and microfracturing of sediments in Lake Annecy in northwestern Alps, implying a brittle-like behavior of soft, water-saturated, sediment) that they interpret as earthquake induced. Sediment gravity flows in the same lake were also interpreted as likely of seismic origin based on (1) the lack of corresponding sub-aerial landslides and (2) the grain sizes being larger than those found in fluvial input aprons, despite the selection of their drilling sites to avoid fluvial input.

In another alpine lake, Schnellmann et al. (2002) interpret five paleoseismic events in the past 15 ka from the evidence of a series of slump deposits in the subsurface of Lake Lucerne. This study identified a stratigraphic “fingerprint” for the sediment deposit associated with the well-described AD 1601 earthquake (Figure 2B.23). The earthquake triggered numerous synchronous slumps and megaturbidites within different sub-basins of the lake, producing a characteristic seismic-stratigraphic linkage between sites imaged with seismic profiling. Four prehistoric events were dated with $^{14}$C measurements and tephrachronology on core samples, and used to establish the recurrence period of similar earthquakes, as well as possible tsunami events through the Holocene.
thickness of the sediment affected by sliding

- <5 m
- 5 - 10 m
- >15 m

- no seismic penetration
- megaturbidite
- slide direction

Fig. 2A

Fig. 2B

seismic survey grid

historic seismic event (A.D.1601)

prehistoric seismic event (2,420 cal. yr B.P.)

prehistoric event (3,240 cal. yr B.P.)

prehistoric event (3,240 cal. yr B.P.)
Other lake records have been described by Carillo et al. (2008) in Venezuela, in Lago Icalma, Chile (Bertrand et al., 2008), in Lake Le Bourget in the NW Alps, (Chapron et al., 1999) in Lake Bramant, western French Alps (Guyard et al., 2007), and other localities.

Niemi and Hall (1994) reported evidence for an association between the 1927 Dead Sea earthquake, a submarine slide in the lake bottom, and a ~1 m tsunami apparently generated during this event. Seismic reflection profiling imaged a shallow slide of broad areal extent involving the particularly stable lake sediments. The lack of other potential triggers for this large slide in part was used to infer earthquake origin, along with the eye witness reports of the tsunami originating in the center of the lake, as opposed to a seiche. Eight other similar slides in the same location suggested a repeat time of significant earthquakes on the Dead Sea Fault of several thousand years over the last 20–30 ka. Subsequently, Marco et al. (1996) investigated a shallow water record from the Dead Sea Graben, formerly lacustrine, and identified a series of seismic disturbances consisting of pulverized laminae, some in association with fault scarps, indicating earthquakes of $M_w > 5.5$. The recurrence interval from this ~40,000 year lacustrine record is ~1600 years, similar to the Dead Sea record. Migowski et al. (2004) report that further investigation of the Dead Sea sedimentary record using varve counting shows that all recent and historical strong local earthquakes could be identified, including the major earthquakes of AD 1927, 1837, 1212, 1033, 749, and 31 BC. A total of 53 seismites were recognized in this study, which also identified long-term patterns of quiescence and greater activity.

Coastal Lakes  Tsunami overwash deposits are well known now from several settings including Sweden, Japan, Kamchatka, and Cascadia. They offer long records and continuous sedimentation, much like deeper marine environments, but are more accessible. Deposits in these settings may be mixed with other events, and thus may present some ambiguities, however the same can be said of any off-fault paleoseismology or tsunami deposit.

Bradley Lake, Cascadia Margin  Bradley Lake, located close to the coastal dunes of Oregon along the southern Cascadia margin, records local tsunamis and seismic shaking on the Cascadia megathrust (Kelsey et al., 2005). The lake stratigraphy includes 13 landward thinning sand sheets interpreted to be tsunami overwash into the lake on the basis of microfossil analysis. The marine incursions included marked changes in salinity of the freshwater lake. Four additional sediment may represent localized turbidity currents from earthquake shaking. The marine incursions had to travel overland to enter the lake, and thus represent a sensitivity test of the magnitude of these tsunamis (Figure 2B.24), which had to be at least 5–8 m above sea level with a duration of at least 10 min. The analysis of the disturbance events in Bradley Lake is analogous to that used for offshore turbidites in that the investigators developed age models for their lake cores, examined sand layers for evidence of basal erosion based on missing section, and uniquely, were able to use the brackish episodes which resulted in varves to establish the interseismic sedimentation rates. The interseismic depositional units included massive muds, but could be used assuming

Figure 2B.23: Slide deposits in Lake Lucerne, Switzerland mapped with high-resolution reflection profiling. Slide deposits related to specific horizons. (A) Grid of 3.5 kHz seismic profiles acquired for this study. (B–D) Distribution and thickness of slide bodies corresponding to three event horizons identified in the reflection profiles. Hachured areas mark extent of megaturbidites directly overlying slide bodies. Bathymetric contour interval is 10 m. From Schnellmann et al. (2002) their Figure 3. Reprinted with permission of the Geological Society of America. (See Color Insert.)
Before tsunami

Tsunami starts to flow into the lake: Erosion of the floor sediments

Tsunami inundates the lake shore: Erosion of peat and vegetation

Max. tsunami water level: Settling of silt and organic particles

Outflow of tsunami: Erosion and redeposition

After tsunami: Sedimentation of fines

Figure 2B.24: Erosion and depositional model for a tsunami inundating a shallow basin on the sea floor and a lake: (A) before tsunami inundation, normal sedimentation. (B) the tsunami inundates and erodes the shore and flows into the lake where it rips up clasts from the lake floor. Sand is deposited in the marine basin (facies 4 and 5). (C) the tsunami inundates and erodes peat and
the same rates found from the varve-based section, establishing interseismic intervals independent of the radiocarbon data. From these varve-based rates, Kelsey et al. (2005) established a sequence of event ages which closely matched the ages based on $^{14}$C. Goldfinger et al. (2009) conclude that the offshore turbidite record closely matches the Bradley Lake disturbances record for the past ~4600 years, and that both sites have recorded both long and segmented ruptures.

Similarly, stratigraphic, geochemical, and microfossil data were collected from sediments in Laguna Mitla, the Pacific coast of Guerrero, Mexico. The rapid relative sea-level rise, marine inundation, and possible tsunami deposit have been interpreted as evidence of a megathrust earthquake and associated tsunami deposits (Ramirez-Herrera et al., 2007).

2B.4.4.3 The Storegga Tsunami

In an example superficially similar to Bradley Lake, tsunami deposits have been found in coastal lakes adjacent to the Storegga submarine slide off western Norway. Bondevik et al. (1997) report distinctive deposits found in small coastal lakes along the Norwegian coast. These lakes, situated from 0 to 11 m above the 7000 year BP shoreline, were sampled with a piston corer and contain a deposit very distinct from the lake sediments in the cores. The base of the deposit is marked by an erosional unconformity which can be correlated around the lake basins. The distinctive normally graded sand overlies the erosive surface, which shows greater erosion toward the seaward side of the basins. Locally, the sand contains shell fragments and foraminifera. The tsunami deposit is a fining upward sequence with occasional massive sand at the base, which includes the marine fossils. The sand thins and decreases in grain size landward direction. Above the fining upward sand, the sequence includes a coarse organic layer with rip-up clasts. The tsunami unit generally fines and thins upward. One of the most convincing pieces of evidence from this unusual tsunami setting is that the basins show stratigraphic thinning by their elevation from the coast (Figure 2B.24). The basins closest to the paleoshoreline ~7000 BP have several sand pulses separated by organic debris, while successively higher basins (6–11 m above the 7000 year shoreline) have only one sandy unit.

Several basins were investigated that were below the 7000 BP paleoshoreline, but that are now exposed due to postglacial crustal rebound. Bondevik et al. (1997) interpret the presence of the tsunami deposit in these basins as well. The character of the deposit in the sub-sea is graded sand beds with occasional organic rich facies between the sand beds. The Norwegian coastal tsunami deposits are linked temporally to the coeval Storegga slide, making this a classic example.

2B.4.4.4 Nankai and Suruga Troughs, Japan

Tsunami deposits have been found along the Japanese coasts in the Nankai and Suruga Trough areas of eastern Honshu. On land, deposits were discovered first at archeological sites, but many of the best vegetation at the lake shore. Sand brought in by the tsunami is deposited in the lake basin (facies 4 and 5). (D) suspended material such as rip-up clasts, twigs, gyttja, sand and silt settles producing normal graded organic beds (facies 6 followed by facies 7). (E) withdrawal of the wave, erosion and redeposition of the tsunami deposits, organic material carried out of the lake. (F) after the tsunami, deposition of suspended fines in addition to organic matter from reworking of tsunami sediments deposited above the lake. Stages (B)–(E) represent the inundation and withdrawal of one tsunami wave. Basins closer to sea level experienced several waves as is shown by the alternation of sand and organic beds. From Bondevik et al. (1997). Reprinted by permission of Wiley Interscience.
deposits are found in coastal lakes (Okamura et al., 2000; Nanayama et al., 2002; Tsuji et al., 2002). Similarly, historical lake tsunami deposits are also found along the Sagami Trough (Fujiwara et al., 2000). Typically these deposits are found in limited sites in the coastal plains and lakes, and do not define landward thinning sheets (Komatsubara and Fujiwara, 2006) as is commonly reported. These deposits resemble the Bradley Lake deposits in that they are typically sandy, fining upward deposits, intercalated with muddy lake sediments, they contain marine fossils including mollusks, nanno plankton, foraminifera, and ostracods (Komatsubara and Fujiwara, 2006 and references therein). These deposits range in thickness from a few centimeter to over 6 m and are sometimes covered with plant fragments. These deposits mostly do not have detailed grain size and stratigraphic analyses, though may correlate to historically recorded earthquakes and in other cases, are dated to within reasonable temporal correlation with historical earthquakes (Komatsubara and Fujiwara, 2006).

2B.4.5 Submarine Landslides Triggered by Earthquakes

Several studies have confirmed the triggering of large submarine landslides and turbidity currents by earthquakes, and numerous others in the geologic record may have been as well. Perissoratis and others (1984) documented a slump covering 15–20 km² in the eastern Korintiakos Gulf along the coast of Greece triggered by a series of earthquakes (Mw = 6.4–6.7) from 24 February to 4 March 1981. Field and others (1982) documented a sediment flow/lateral spread on a 0.25° slope on the submarine Klamath River delta off the coast of northern California; the feature extends along 20 km of the delta front and is about 1 km long (from scarp to toe). The very low slope and the presence of liquefaction features on the surface both suggest seismic triggering, and repeated bottom surveys before and after the Mw = 6.5–7.2 offshore earthquake of 8 November 1980 conclusively linked the landslide to the earthquake. Lee and Edwards (1986) analyzed the stability of four submarine landslides off the coasts of California and Alaska and concluded that three of them required seismic shaking to have triggered failure.

These studies provide the basis for interpreting older submarine landslide deposits in terms of seismic triggering. Examples of other submarine slides and mass wasting deposits include Viscaino et al. (2006), who report on turbidites and a submarine slide in the Marqués de Pombal area of the Iberian margin. They report a mixed record of slides and turbidites in which a large landslide deposit observed in acoustic backscatter imagery is not related to the 1755 Lisbon Earthquake, but is much older, with ages between ca. 3270 and 1940 yr BP. They found that the deposit more likely related to the 1755 Lisbon earthquake is a thin turbidite.

The majority of historical submarine landslides have been linked to earthquakes, including the well-known Grand Banks earthquake of 1929 which spawned a landslide and associated turbidity current that broke a series of submarine cables downslope, thus recording the direction and speed of travel (Heezen and Ewing, 1952). A near repeat of this event occurred on 26 December 2006, when the magnitude 7.1 Hengchun earthquake was followed by the breakage of eleven submarine cables in the Strait of Luzon, between Taiwan and the Philippines (Hsu et al., 2008).

The causes of submarine landslides may be complex, beginning with a weak depositional sequence that may be climate related, further weakening from a secondary effect such as gas hydrate destabilization, and then final triggering by an earthquake (Masson et al., 2006). Linkages to earthquake triggers for prehistorical events can be problematic as these factors may be difficult to disentangle. The Storegga slide is thought to be just such an event (Bryna et al., 2005). Submarine landslides may occur coincidentally with earthquakes and may also be responsible for enhancing tsunami generation locally, while at the same time making seismologic interpretation difficult, as in the 1998 Papua New Guinea earthquake/tsunami
(Satake and Tanioka, 2003). While equivocal, this and perhaps other similar events may represent slumps where rapid rotation and seafloor offset generate significant seafloor motion and tsunami from a modest earthquake source (Matsumoto and Tappin, 2003).

Numerous other submarine slides have likely earthquake origins, but have not been documented to the level required to establish this linkage. Among the largest known are the super scale slides of the Oregon margin. Using SeaBeam bathymetry and multichannel seismic reflection records on the southern Oregon continental margin, Goldfinger et al. (2000) identified three large submarine landslides involving ~8000 km$^2$, and a volume of ~12,000–16,000 km$^3$ of the accretionary wedge. The three arcuate slump escarpments are nearly coincident with the continental shelf edge on their landward margins, spanning the full width of the accretionary wedge. Debris from the slides is buried or partially buried beneath the abyssal plain. The ages of the three major slides decrease from south to north, indicated by the progressive northward shallowing of buried debris packages, increasing sharpness of morphologic expression, and southward increase in postslide reformation of the accretionary wedge. The ages of the events, derived from calculated sedimentation rates in overlying Pleistocene sediments, are approximately 110, 450, and 1210 ka. This series of slides traveled 25–70 km onto the abyssal plain in at least three probably catastrophic events, which may have been triggered by subduction earthquakes. The slides would have generated large tsunami in the Pacific basin, possibly much larger than that generated by an earthquake alone. The authors also identified a potential future slide locality with incipient breakaway features off southern Oregon that may be released in a subduction earthquake.

2B.4.6 Coeval Fault Motion and Fluid Venting Evidence

Offshore faults leave evidence of movement and timing in the form of scarps, colluvial wedges, liquefaction, and mass wasting deposits similar to onshore faults, but also in the form of fluid venting. Seafloor evidence of fluid venting is commonly expressed as chemical fluxes (e.g., Gamo et al., 2007) and occasionally as carbonates originating in the commonly methane rich fluids. Carbonates are typically observed as ribbons in fractures, and as “chimneys” composed of annular venting laminae with a vertical orientation (e.g., Goldfinger et al., 1996; Ogawa et al., 1996). ROV and submersible studies, as well as experimental results from flow meters and borehole CORKs (Circulation Obviation Retrofit kit) in ODP drill holes, suggest such fluid flow is invigorated in the rupture region following a significant earthquake, thus making it direct evidence of paleoearthquakes. For example, recent observations at the Nankai Trough Sites 808 and 1173 (Figure 2B.6) illustrate the capture of strain via formation pressure in hydrologically isolated sediment sections. Two deformational events have been captured, the first contemporaneous with a very low-frequency earthquake swarm in the prism, and the second contemporaneous with an earthquake swarm in the subducting plate. The transients reflect both coseismic strain and postseismic relaxation. These are superimposed on a rise in pressure due to local interseismic strain accumulation. Similar observations have been made at other fault sites in Cascadia and Costa Rica using both flow meters and CORKs.

Okamura et al. (2005) report the rupture extents of paleoearthquakes in the Sea of Japan can be defined by the extent of scarp activity and fluid expulsion along the fault trace using submersible observations. Quaternary folds and fault zones here comprise several arc-parallel zones, known as the Okushiri Ridge, Sado Ridge, Awashima to Oga ridge, and several others. Historical earthquakes occurred in 1940, 1983, and 1992 along the Okushiri Ridge. Shinkai 6500 submersible dives in the source area of the recent earthquakes widely observed fresh fissures, slope failures, venting and debris on the slope above active segments in the areas of recent earthquakes, and fissures, where similar features were covered with muddy
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sediments in areas between these ruptures. Evidence of submarine fluid and gas venting directly related to
earthquakes has been observed with the Izmit 1999 event (Kuscu et al., 2005) and other localities.
Kitamura et al. (2002) report a unique sub-aqueous sand blow deposit associated with the 1995 Kobe
earthquake.

Acknowledgments

My thanks to Takeshi Nakajima for his review of this chapter, and to Gordon Seitz, John Hughes Clarke, Hannes Grobe, Earl Davis,
Celia McHugh, Takeshi Nakajima, Tom Rockwell, Mark Legg, Waldo Wakefield, and Klein-L3 for generous assistance with and
permission to use data and figures from their work.