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Coastal erosion and recovery from a Cascadia subduction zone earthquake and tsunami

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Abstract

Tsunamis generated by great earthquakes threaten coastal infrastructure, development, and human life. Earlier work has documented the inland extent and frequency of past tsunamis, but little is known about the magnitude of material eroded during prehistoric tsunamis and how erosion and coastal recovery are recorded in coastal stratigraphy. In this study we use high-resolution ground-penetrating radar (GPR) to image and quantify coastal erosion experienced during a late Holocene (~900 cal BP) Cascadia subduction zone earthquake and tsunami along the northern California coast. The GPR profiles illustrate three stratigraphic signatures created during co-seismic subsidence and tsunami erosion and coastal recovery. The first is erosional truncation of the underlying seaward-dipping reflections created by pre-tsunami normal beach progradation. The second is a series of landward-dipping, flat-lying, and channelized reflections marking the filling of erosional topography and coastal reworking of the irregular shoreline following inundation and erosion. The third is an abrupt landward termination of the stratigraphic unit marking coastal straightening and post-tsunami rejuvenation of normal coastal progradation. Erosion from the ~900 cal BP earthquake and tsunami extended more than 110 m inland of the contemporary shoreline and removed/reworked 225,000 ± 28,000 m^3 of sand from a 1.7-km stretch of the coast, far exceeding anything experienced during historical El Niños along the Pacific Coast of North America. This study provides the first quantitative estimate of the amount of coastal erosion from a pre-historic earthquake and tsunami and outlines a strategy for estimating erosion during similar events elsewhere.

1. Introduction

Recent tsunamis created by large megathrust earthquakes in Sumatra, Chile and Japan resulted in large losses of life and extensive damage to coastal infrastructure (Bondevik, 2008; Goto et al., 2011; Mimura et al., 2011). These events have renewed efforts to understand the impacts and frequency of past tsunamis on coastlines. New studies have improved our ability to identify their deposits in the sedimentary record (Gelfenbaum and Jaffe, 2003; Switzer et al., 2006; Morton et al., 2007; Gouramanis et al., 2015) and determine the inland extents and frequencies of past tsunamis across the globe (Kelsey et al., 2002; Schlichting and Peterson, 2006; MacInnes et al., 2016). However, despite these advances, identifying past tsunami deposits in areas marked by sandy shorelines without muddy or peaty back-barrier deposits remains difficult.

An important impact of tsunamis is coastal erosion (Paris et al., 2009). Recent tsunamis in Sumatra (Paris et al., 2009; Liew et al., 2010) and Chile (Morton et al., 2011) resulted in extensive erosion of low-lying coastal regions. Coastal erosion can remove important natural resources needed not only for local economic extraction but also as a magnet for the increasingly important tourism industry along many coastal regions (Adger et al., 2005). However, few methods have been developed to quantify coastal erosion experienced during past earthquakes and tsunamis. Understanding the nature of past coastal erosion experienced during earthquakes and tsunamis could be an important tool for establishing proper setbacks and foundation designs for the development of coastal infrastructure.

Tsunamis created by late Holocene ruptures of the Cascadia subduction zone (CSZ) inundated much of the northern California coast (Peterson et al., 2011; Valentine et al., 2012; Fig. 1). The sandy beaches of the coastal plain north of Crescent City, California provide an excellent natural laboratory to explore the impacts of past earthquakes and tsunamis on sandy shorelines. In this study we use ground-penetrating radar (GPR) and optically stimulated luminescence (OSL) dating to build a model of how co-seismic subsidence and tsunami impacts are recorded in beach stratigraphic sections and provide the first quantitative estimates of coastal erosion from prehistoric co-seismic subsidence and tsunami erosion.

2. Study Area

The CSZ marks subduction of the Juan de Fuca plate beneath the North American plate. It stretches for over ~1000-km across the western Pacific margin of North America (Fluck et al., 1997; Fig. 1). Great (>Mw 8) earthquakes, resulting from ruptures within the fault zone, occur approximately every 300-500 years (Atwater, 1987; Atwater and Hemphill, 1987; Nelson et al., 2006; Goldfinger et al., 2012; Milker et al., 2016). These earthquakes are known to have generated not only coastal subsidence (Atwater, 1987; Shennan et al., 1996; Nelson et al., 2008; Hawkes et al., 2011) but also great tsunamis (Kelsey et al., 2005) that struck the coast from Vancouver Island, British Columbia (Clague et al., 2000) to northern California (Clarke and Carver, 1992; Valentine et al., 2012). The most recent tsunamigenic event occurred January 26, 1700 (Satake et al., 1996) with two preceding events around 800 cal BP (Nelson et al., 2008; Schlichting and Peterson, 2006) and 1000 cal BP (Kelsey et al., 2005; Schlichting and Peterson, 2006). The turbidite record of events also suggests another great earthquake occurred around 500 cal BP with potentially three other smaller events after 1000 cal BP (Goldfinger et al., 2012), although coastal records for these events have yet to be identified (Nelson et al., 2006; Milker et al., 2016). Their absence in the coastal record may be due to not being large enough to create a signature or creating a signature too small for preservation in coastal settings.

The Crescent City coastal plain is a low-elevation (<30 m) expanse of Quaternary Smith River alluvial, late Pleistocene marine terrace, and Holocene strandplain and aeolian deposits located between the Saint George and Smith River Faults within the Lake Earl syncline (Polenz and Kelsey, 1999). The southwest-northeast oriented contraction is a result of overall convergence along the CSZ (Polenz and Kelsey, 1999). Subsidence within the syncline led to the development of a relatively flat, low-lying coastal plain. Sediment delivered by the Smith River at the northern portion of the coastal plain is generally transported south forming a prograding strandplain marked by sandy beach and dune deposits backed by a coastal lake, Lake Earl (Fig. 1; Polenz and Kelsey, 1999). These Holocene coastal deposits are backed by late Pleistocene marine terraces stepping up to the abruptly rising Klamath Mountains (Fig. 1). The area is particularly susceptible to tsunamis not only generated by the CSZ but also within the Gulf of Alaska (Peterson et al., 2011). A record of no fewer than six past tsunamis are preserved in marshes from the southern portions of the Crescent City coastal plain (Peterson et al., 2011; Fig. 1). The two most recent of which are dated to 270-560 cal BP and 784-954 cal BP and are thought to represent CSZ earthquakes and tsunamis (Peterson et al., 2011).

3. Methods

3.1 Ground-penetrating radar (GPR)

We collected approximately 20 km of GPR data using a Sensors and Software EkkoPulse Pro. Initially lines were collected with 100, 200, and 500 MHz antennae. Common-midpoint depth surveys were conducted at two locations within the coastal strandplain to determine the velocity of the radar waves within the subsurface. GPR profiles were topographically corrected using RTK-GPS survey data collected at the same time as the GPR. All elevation data was corrected to NAVD88 using the online OPUS website (www.ngs.noaa.gov/OPUS/; last accessed August 2016) and reported as elevation above NAVD88 mean sea level. Processing of the GPR lines included automatic gain control and DeWow (a proprietary Sensors and Software processing algorithm). GPR profiles were loaded into EkkoView Deluxe software and ArcMap for interpretation.

Four vibracores were collected in 7.2-cm diameter aluminum tubes in order to obtain material for OSL dating and ground-truth GPR interpretations (Fig. 1). The core locations were picked to sample deposits landward, seaward, and within a GPR unit thought to represent erosion from a CSZ earthquake and tsunami. After removing sections for OSL, the cores were split, photographed, and described at the sedimentology lab at the University of California Santa Barbara. Facies within the cores were distinguished based on sorting and sedimentary structures.

3.2 Optically stimulated luminescence (OSL)

Optically stimulated luminescence measures the time since sand or crystal grains were last exposed to sunlight. Bøtter-Jensen et al. (2003), a review by Rhodes (2011), and a series of dedicated contributions in the recently published Encyclopedia of Scientific Dating Methods (Rink and Thompson, 2013) give a detailed description of the OSL method. The method works well for dating sediments from depositional settings where exposure to sunlight is common such as in aeolian (Ballarini et al., 2003;

Bateman, 2008; Rhodes, 2011) and beach deposits (Murray et al., 1995; Murray and Funder, 2003; Rhodes, 2011; Simms et al., 2012) but can be problematic when attempting to date fully marine and fluvial deposits (Olley et al., 1998; Wallinga, 2002; Rhodes, 2011). A major challenge with using OSL in fully marine and fluvial deposits is the possibility that the grains were not fully bleached, i.e., received enough sunlight to “reset” the OSL signal. However, insufficient bleaching is usually not the case within aeolian and beach deposits (Murray et al., 1995; Murray and Funder, 2003). In studies in which independent age constraints are available (e.g. ^{14}C, known last interglacial shoreline), OSL ages on beach deposits are typically within 10% of the expected age (Murray and Funder, 2003; Simms et al., 2012). Other possible sources of error affecting the samples in this study are variable water content during the depositional period and intake or leaching of radioactive nuclides.

Five ~25-30 cm sections of the unsplit cores were cut from the vibracores for OSL dating as to avoid exposure to light prior to dating. This much section was needed to assure obtaining enough quartz for OSL dating. Given the nature of beach sedimentation (prograding laterally rather than aggrading vertically) sampling over such a large vertical range within the cores likely contributes little to age uncertainties. Modern rates of beach progradation average 0.7 m/yr in the region (Hapke et al., 2006) and 5 measured beach profiles yielded beach gradients ranging from 0.06 and 0.19 with an average of 0.14. Thus the 25-30 cm vertical sections of the cores were likely deposited within 2-5 years.

Samples for OSL were prepared in a dark room under subdued red-light conditions. The light exposed ends were removed from the cores (approx. 8 cm from each side of the core; >500g), dried and used for gamma spectrometry. From the remaining material (10-15 cm in the center of the core) quartz separates were prepared by treating 90-150 µm or 150-180 µm diameter grains with 27% H_2O_2, 10% HCl, and 48% HF for 40 min, and subsequent density separation with lithium polytungstate solution (densities 2.75 g/cm^3 and 2.62 g/cm^3). We used 3mm quartz aliquots, prepared on stainless steel discs using silicone spray. Measurements were conducted using a Risø TL/OSL-DA-20 reader, Risø National Laboratory, with a bialkali PM tube (Thorn EMI 9635QB) and Hoya U-340 filters (290-370 nm). The built-in $^{90}Sr/^{90}Y$ beta source gives a dose rate of 120 mGy/s. Optical stimulation was carried out with blue LEDs (470 nm), delivering 82 mW/cm^2 to the sample. IR stimulation was from an IR LED array at 875±80 nm with 124 mW/cm^2 power at the sample. The heating rate used was 5 °C/s.

We used a Single-Aliquot Regenerative-dose (SAR) method with high temperature bleach for determination of the equivalent dose (Wintle and Murray, 2006). Dose recovery and plateau tests resulted in a preheat temperature of 200°C. Dose responses were fitted with a linear function. Samples generally showed low signals and bad recycling ratios, so that large error margins for the selection criteria had to be chosen. Aliquots with recycling ratios between 0.8 and 1.2, dose recovery better than 20%, and IR depletion <15% were used for calculating the equivalent dose (Wintle and Murray, 2006; Duller, 2003) based on the common age model (Galbraith, 1999).

U, Th, and K concentrations in the samples were measured with high resolution Ge gamma spectrometry with a Reverse Electrode Coaxial Germanium detector from Canberra Industries, Inc. No disequilibria in the uranium decay chain were observed. Water content (mass of water divided by mass of dry sample) was measured as preserved in the cores taken. To allow for minor losses of water during the coring process we added 0.03 to the measured water content and also used 0.03 as uncertainty for the measured values. Dose rate from cosmic rays was determined from the depth of sample below the surface along with its longitude, latitude and altitude, as described by Prescott and Hutton (1994).

3.3 Calculation of erosion

We used the difference in elevations between an erosional surface imaged in the GPR and the modern surface to estimate the amount of coastal material eroded and/or reworked during the erosional event that created the surface. We assumed that the seaward-dipping reflections imaged below the erosional surface continued up to the ground surface (Fig. 2). We determined a volume of sediment removed by averaging the cross-sectional area of erosion within the two GPR profiles ($EA_{GPR\text{-}line18}$, $EA_{GPR\text{-}line01}$) oriented perpendicular to the coast (a third line images part of the erosional surface but the erosion continues landward of the line and thus only provides a minimum estimate) that contained the erosional surface multiplied by the length (L_{er}) in which the erosional surface could be traced:

$$V = ((EA_{GPR\text{-}line18}+EA_{GPR\text{-}line01})/2) \times L_{er}$$

We assumed an area measurement error (ε_{am}; uncertainty in GPR velocity, survey errors) of 10%, encompassing the variability in velocity estimates (8%) plus the percent difference of the GPR-recorded length and GPS-measured length (~2%). When combined with one half of the difference between the two area calculations (Δ_a) gave a total area error (ε_a) derived by the following expression:

$$\varepsilon_a = ((0.5 \times \Delta_a)^2 + \varepsilon_{am}^2)^{0.5}$$

This error was combined with a length measurement error (ε_L) of 2 m, the uncertainty in the distance between the two lines, to determine the total volumetric error (ε_v) using the following expression:

$$\varepsilon_v = (V \times A_{avg}) \times ((\varepsilon_a / A_{avg})^2 + (\varepsilon_L / L_{er})^2)^{0.5}$$

4. Results

GPR profiles through the Crescent City coastal plain contain three GPR facies assemblages comprised of 2-3 GPR facies each. The first, GPR facies assemblage A, consists of three genetically-related facies. These include a landward-dipping (GPR facies A_{ld}), flat (GPR facies A_f), and concave-up (GPR facies A_c) series of parallel to subparallel reflections (Figs. 3, 4). The second GPR facies assemblage, GPR facies assemblage B, contains 2 GPR facies and is generally found beneath reflections of GPR facies assemblage A (Fig. 3). GPR facies B1, within GPR facies assemblage B, consists of a series of seaward-dipping, parallel reflections (Fig. 3). GPR facies B1 is only found in GPR profiles collected between dune ridges and most prevalent in the low-lying region in the middle of the Crescent City coastal plain seaward of Lake Earl (Fig. 1). The second GPR facies within GPR facies assemblage B is GPR facies B2 (Figs. 4, 5). It consists of parallel, wavy, subhorizontal reflections and is found near the ephemeral inlet connecting Lake Earl with the open ocean. Overlying reflections of GPR facies assemblage A downlap on top of GPR facies B2. GPR facies assemblage C consists of an assortment of landward-dipping (GPR facies C_{ld}), flat-lying (GPR facies C_f), and channelized reflections (GPR facies C_{ch}) (Fig. 4). It is only found above an erosional surface cut into GPR facies B1 (see below) but stratigraphically below GPR facies assemblage A (Fig. 2).

An exceptionally well-developed succession of GPR facies B1 is found within the central portion of the study area in a region between the first two large discontinuous rows of dunes (Fig. 2). This succession is interrupted by an erosional surface, ES_1, with relief up to 2.5 m in amplitude (Fig. 2). The erosional surface truncates underlying seaward-dipping reflections (GPR facies B1) and shallows landward (Fig. 2). It is overlain by reflections of GPR facies assemblage C (Fig. 2). The erosional surface is terminated at its seaward end by a sharp seaward-dipping surface that marks continuation of the undisturbed seaward-dipping reflections of GPR facies B1 (Fig. 2). It extends a minimum of 110 m inland and cuts to a depth of 1.5 m below the pre-erosional ground surface (2.5 m below the post-erosional ground surface; Fig. 2). The surface is imaged in 3 consecutive shore-perpendicular profiles over a distance of 1.7 km (Figs. 1, 2). To the north, the hosting GPR facies B1 as well as the overlying GPR facies assemblage C are buried beneath GPR facies assemblage A to a depth greater than the penetration of the GPR signal and erosional surface ES_1 is no longer mappable. To the south, the hosting GPR facies B1 and the erosional surface are replaced by a section of GPR facies assemblage A that downlaps onto a horizontal surface below which lies GPR facies B2 (Fig. 5). This section of GPR facies B2 overlies and potentially onlaps, another erosional surface, ES_2, which truncates horizontal or possibly landward-dipping parallel reflections (Fig. 5). The amount of material above erosional surface ES_1 along the 1.7 km stretch of coastline that the surface can be mapped in GPR (3 shore-parallel lines; Fig. 1) is 225,000 ± 28,000 m^3.

The vibracores reached depths up to 210 cm and contain two major sedimentary facies (Fig. 6). The first (S_l) is a moderately to well-sorted laminated dark gray medium to coarse sand with occasional pebble lenses. This sedimentary facies is found in cores TD15-02, TD15-03 and TD15-04 (Fig. 6). The second sedimentary facies (S_{ps}) is a poorly sorted pebbly black to dark grayish brown medium to coarse sand. Sedimentary facies S_{ps} is generally darker in color (2.5Y 3/2) than sedimentary facies S_l (2.5 Y 4/1) and is found within core TD15-01 (Fig. 6). Both sedimentary facies contain roots and plant fragments although some of the woody material in sedimentary facies S_l appears to be abraded.

Five OSL ages were obtained from the vibracores. The OSL ages from cores TD15-03 and TD15-04 as well as the lowest sample from TD15-02 were obtained from sedimentary facies S_l and GPR facies B1, while the shallower sample from TD15-02 was obtained from sedimentary facies S_{ps} and GPR facies C_{ch} (Fig. 2). The sample from core TD15-01 was from sedimentary facies S_{ps} and was taken landward of the GPR line. Results for the OSL samples are listed in Table 1. Overdispersion values (Galbraith et al., 1999) have been calculated to evaluate incomplete resetting of the samples prior to deposition. The values would also be increased, if the extended vertical range of the OSL material represented a large age range. Overdispersion values range from 4-15%, which is well within the range for well bleached samples (Arnold and Roberts, 2009), indicating sufficient resetting of the samples prior to deposition and a negligible influence of the core thickness. The ages range from 807 ± 49 years to 1036 ± 52 years (Figs. 2 and 6; Table 1, 1-σ error limits). The two ages from the same core (TD15-02) are in stratigraphic order and bracket the age of erosional surface ES_1 to between 929 ± 44 and 1036 ± 52.

5. Discussion

5.1 GPR facies interpretations

GPR facies assemblage A is interpreted to represent aeolian dune deposits forming a veneer of sediments across most of the seaward portions of the Crescent City coastal plain. This interpretation is based on its location within modern dune ridges of the coastal plain and similarity with GPR facies described in similar coastal areas (Tamura et al., 2011). Based on its seaward-dipping character, stratigraphic position beneath deposits interpreted as dunes (GPR facies assemblage A), and similarity to other GPR profiles collected along prograding coasts (e.g. Rodriguez and Meyers, 2006; Hein et al., 2013), GPR facies B1 is interpreted as prograding beach deposits. GPR facies B2 is interpreted as tidal inlet deposits of the tidal inlet connecting Lake Earl with the open ocean. This interpretation is largely based on the location of this GPR facies immediately adjacent to the modern tidal inlet. The surface bounding GPR facies B2, ES_2, is thus interpreted as a tidal ravinement surface caused by the migration of the tidal inlet.

5.2 Origin of erosional surface ES_1

Possible causes of erosional surface ES_1 include fluvial erosion from the Smith River, a tidal inlet connecting Lake Earl with the open ocean, or a marine-sourced erosional event such as a tsunami or large storm. Abandoned meanders of the Smith River are well-developed along the northern portions of the Crescent City coastal plain suggesting widespread migration of the river (Fig. 7). The best preserved and likely youngest of these abandoned meanders south of the current active channel is Yontocket Slough (Fig. 7). Prior palynological work constrained by bulk sediment radiocarbon ages suggest the slough was abandoned prior to 926-1405 calendar years BP (Bicknell and Austin, 1991; calibrated using Calib 7.1; Reimer et al., 2013). Any younger paleo-channels of the Smith River would likely also be preserved. The absence of better preserved features south of Yontocket Slough suggests the Smith River was >5 km north of the imaged primary erosion at the time of its formation, ~ 900 cal yrs BP. In addition, the Smith River erodes to depths below ground surface of >4.5 m (Parish and Garwood, 2016), well below the lower elevations of erosional surface ES_1. A tidal inlet origin is also unlikely due to the shore-parallel nature of the erosional surface and its different expression than that seen in GPR lines near the modern inlet (ES_2; Fig. 5). A marine-source for the process(es) responsible for the formation of ES_1 erosion is supported by the seaward deepening nature of the erosional surface, its alignment parallel to the coast, and its similar geometry as other GPR-imaged marine erosion surfaces (Meyers et al., 1996; Buynevich et al., 2007).

While all but the youngest and oldest OSL ages overlap within error, the ages from the core that intersects the erosional surface itself bracket the surface to between 1036 ± 52 and 929 ± 44 years. This age overlaps within error of the penultimate CSZ great earthquake and tsunami. The event is dated locally at 784-954 cal BP (Peterson et al., 2011) and elsewhere between 800-1000 cal BP (Kelsey et al., 2005; Schlichting and Peterson, 2006; Nelson et al., 2008; Goldfinger et al., 2012). Based on its marine-source characteristics and overlapping age with a known Cascadia subduction zone earthquake, we interpret erosional surface ES_1 to be the product of the ~900 cal BP CSZ earthquake and tsunami.

The 807 ± 49 years age in core TD15-03 within GPR facies B1 seaward of ES_1 fits well with a post-tsunami age for renewed normal shoreline progradation. The unexpectedly young age of 908 ± 48 years in core TD15-01 located landward of the erosional surface could be a result of the depositional facies of the unit. Our GPR line did not extend to the location of that core due to a wet marsh environment and thus we are not certain that it sampled a section of GPR facies B1, prograding beach deposits, or GPR assemblage C, interpreted as post-tsunami coastal deposits. The sediments within this core, sedimentary facies S_{ps}, are poorly sorted and darker in color than those within the other three cores, sharing more sedimentary characteristics with the sediments above ES_1 in neighboring TD15-03 than the underlying and more seaward prograding beach deposits. The fifth age of 957 ± 48 years is out of stratigraphic order with the other ages (assuming a seaward younging of the sediments). The date could represent the age of a lower sand unit (toe of the shoreface of a more landward sea-ward dipping foreshore unit). All samples had comparable OSL properties so that this “outlier” cannot be explained at the present time, without obtaining further samples from the site.

The OSL ages within the prograding GPR facies B1 at the base of core TD15-02 and within core TD15-03 located 75 m to the west suggest a minimum beach progradation rate averaging 37 cm/yr. Hapke et al. (2006) determined historical progradation rates for the Crescent City coastal plain using aerial photographs and coastal surveys since the 1870’s. According to their work, the section of the coast seaward of the imaged erosional surface has been stationary but increases to more than 1 m/yr to the north near the mouth of the Smith River (Hapke et al., 2006; redrawn in our Fig. 1). If the Smith River were located farther south, such as the location of Yontocket Slough, around 900 cal BP, a progradation rate of 37 cm/yr would fall within the range of modern progradation rates at similar distances from the mouth of the Smith River. As the river mouth migrated farther to the north, the rate of progradation at the location of ES_1 likely slowed because of the increasing distance to the sand source (the Smith River). Alternatively, along other subduction-zone margins, large earthquake disturbances are thought to increase the rate of beach progradation by increasing the sediment supplied to the coast due to increased hillslope failures and other disturbances within the watersheds of the surrounding region (Goff and Sugawara, 2014).

Meyers et al. (1996) interpreted similar surfaces in a Washington State barrier/split complex as transgressive erosion caused by co-seismic subsidence during a CSZ earthquake. Although we cannot rule out co-seismic subsidence as a contributor to the erosion, we favor a dominant tsunami origin for the erosional surface. We favor a tsunami origin rather than a transgressive surface due to its undulating nature compared to the smoother, more concave-up surface imaged by Meyers et al. (1996). Co-seismic subsidence this close to the coastline would result in the landward encroachment of the shoreface. Wave erosion on exposed beach faces is very efficient at producing a concave upward profile in sandy beaches (Bruun, 1954). If the surface we imaged were caused predominately by coastal transgression during co-seismic subsidence, it would not likely preserve isolated remnants seaward of the main scarp, which exhibit 2 m of relief in as little as 50 m (Fig. 2). The undulated nature of the surface is similar to the scoured and irregular shoreline documented after large tsunamis in Sumatra, Chile, and the Kuril Islands (Umitsu et al., 2007; Choowong et al., 2009; MacInnes et al., 2009; Liew et al., 2010; Morton et al., 2011).

5.3 Model for tsunami erosion and coastal recovery

Our ground-penetrated radar (GPR) profiles through the Crescent City coastal plain illustrate three stratigraphic signatures (I, II, III) recording changes in the coastline experienced during co-seismic subsidence and/or tsunami erosion and coastal recovery (Fig. 8). Following an initial pre-tsunami phase of coastal progradation (Fig. 8B), the first signature is an erosional surface cutting across prograding beach deposits (GPR facies B1)("I" Fig. 8D). The second signature is a unit of heterogeneous GPR reflections exhibiting channelized, flat-lying, and landward dipping geometries (GPR facies assemblage C; "II" Fig. 8 D) that infills the erosional topography. The third is an abrupt transition back to the seaward-dipping parallel reflections (GPR facies B1)("III" Fig. 8F). We interpret these three phases of erosion/deposition following initial normal beach progradation (Fig. 8A and B) as: 1) the initial erosion by the tsunami with a possible landward encroachment of the ocean due to co-seismic subsidence ("I", Fig. 8C and D), 2) the changing coastal morphology associated with post-syn-tsunami infilling and reorganization of the coast (“II”, Fig. 8C and D), and 3) a phase of continued coastal progradation initiated by abrupt seaward termination of the phase II deposits caused by shoreline straightening after the coastal systems regain equilibrium and continue to prograde (Fig. 8E and F). A similar sequence of events was observed across the Thailand coast following the 2004 Boxing Day tsunami (Choowong et al., 2009) and across Chile in 2010 (Morton et al., 2011).

5.4 Inland extent and volume of sand removed by earthquake and tsunami erosion

During tsunamis of the last two decades, erosion is generally seen as either the development of a shore-parallel scarp (Morton et al., 2011) representing the retreat of the coastline or as scour channels or pits representing obstacle scours or return-flow channels (Morton et al., 2011; Richmond et al., 2012). The erosional surface imaged in our GPR profiles is likely analogous to the formation (and later burying) of the erosion scarp documented along coastlines following historical tsunamis (Morton et al., 2011). The inland extent of erosion (imaged up to 160 m) is similar to that experienced along Sumatran beaches during the 2004 tsunami (150 m; Paris et al., 2009) and along the Japanese coast in 2011 (~180 m; Nakamura et al., 2012). The beach scarp formed near Purema, Chile, following the 2010 tsunami resulted in up to 95 cm of vertical planation along the coastline and formed up to 170 m inland from the pre-tsunami shoreline (Morton et al., 2011). At other sites along the Chilean coast, as much as 2 m of vertical planation was observed at the coastline (Morton et al., 2011), similar to the 1.5 to 2.5 m observed in our GPR profiles from Crescent City. Similarly, following the 2011 Tohoku earthquake and tsunami, some areas of the Misawa coast were eroded to a depth of 50-100 cm (Nakamura et al., 2012).

The 225,000 ± 28,000 m^3 of erosion experienced along this part of the coast during the ~900 BP CSZ earthquake and tsunami exceeds that experienced by the largest historical El Nino’s along the Pacific Coast of North America (Revell et al., 2002; Thornton et al., 2006). For example, this 1.7-km stretch of coast lost nearly the same amount of sand during the ~900 BP CSZ earthquake and tsunami as the entire 11-km Netarts littoral cell of northern Oregon during the 1997-1998 El Nino (250,000 m^3; Revell et al., 2002). Similarly, the volume of sand lost from this section of the Crescent City coastal plain is one-third of that lost from the beaches across the entire ~17-km extent of southern Monterey Bay, California during the 1997-1998 El Nino (773,000 m^3; Thornton et al., 2006). The amount of erosion per unit distance along the coast (133 m^3/m) is of a similar magnitude to that measured from the 2004 Boxing Day tsunami (83 m^3/m; Matsumoto et al., 2010; 80 m^3/m; Paris et al., 2009).

6. Conclusions

A CSZ earthquake and tsunami at ~900 BP left a distinct mark in the beach stratigraphy of the Crescent City coastal plain of northern California. The three phases left within the beach stratigraphy include 1) the formation of a landward shallowing erosional surface, 2) a unit of hetergenous GPR reflections above the erosional surface, and 3) an abrupt transition back to normal beach progradation marked by seaward-dipping reflections. Based on GPR profiles, we estimate that the event removed 225,000 ± 28,000 m^3 of sand from a 1.7-km stretch of the coast. This volume of sand removal exceeds the amount of erosion experienced during any historical El Nino storm across the Pacific Coast of the United States. This study provides the first quantitative estimates of erosion experienced during a prehistoric subduction zone earthquake and tsunami. In addition, it shows that GPR is an excellent tool for quantifying coastal erosion during past earthquakes and tsunamis.

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Figure Captions

Table 1. OSL Ages

Figure 1. Digital elevation model (DEM; Gesch et al., 2002; Left) and aerial photograph (from ESRI; Right) of the study area illustrating the location of the ground-penetrating radar (GPR) profiles and cores (green circles) collected as part of this study. Also shown is the extent (tan polygon) of the tsunami erosional surface discussed in the text. Historical beach erosion and accretion rates are shown as a white line (Hapke et al., 2006). Unless noted by “-“, scale bar boxes for the DEM are the lower bounds of the elevation ranges in meters. BC = British Columbia, WA = Washington, OR = Oregon, CA = California, CSZ = Cascadia subduction zone.

Figure 2. Uninterpreted (A) and interpreted (B) GPR line TD13_10_18 illustrating the remarkably well-preserved GPR expression of tsunami coastal erosion and recovery. Also illustrated are the locations of the vibracores used to obtain material for OSL dating. Black boxes are the portions of the cores used for OSL ages. Uninterpreted (C) and interpreted (D) GPR line TD15-01 illustrating the same erosional surface to the north. See Figure 1 for GPR line locations and Figure 6 and Table 1 for a complete description of the vibracores and OSL ages, respectively.

Figure 3. GPR line TD15_08_19 illustrating GPR facies assemblage A and GPR facies B1 found within the seaward portions of the Crescent City Coastal Plain. See Figure 1 for line location.

study.

Figure 5. Uninterpreted (A) and interpreted (B) GPR line TD15-03 illustrating the surface replacing the erosional surface highlighted in Figure 2. Also note GPR facies B2. See Figure 1 for GPR line location.

Figure 6. Core descriptions. See Figure 1 for core locations.

Figure 7. Aerial photography of the Smith River and other geomorphic features created by the ancestral Smith River. Aerial imagery from ESRI. See Figure 1 for general locations.

Figure 8. Model for the development of GPR characteristics before, during, and after a tsunami. Also shown are satellite images of Lhoknga, Sumatra prior, one month after, and four years after the December 26, 2004 Boxing Day tsunami (from GoogleEarth).