# Tsunamis and tectonic deformation at the northern Cascadia margin: a 3000-year record from Deserted Lake, Vancouver Island, British Columbia, Canada

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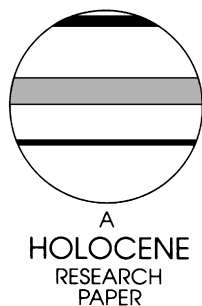
Abstract: Three layers of sand occur within a sequence of muddy gyttja in a 3 m core recovered from Deserted Lake, which lies at the head of a fjord on the central west coast of Vancouver Island, British Columbia, Canada. The sedimentology and microfossil assemblages of these layers are consistent with deposition by tsunamis. Radiocarbon ages suggest that the tsunamis were generated by plate-boundary earthquakes at the Cascadia subduction zone about 2600, 1600 and 300 years ago. Diatom and protozoan contents of the gyttja layers indicate that relative sea level fell in the periods between tsunamis. The sea-level changes are probably a product of interseismic uplift of the North America plate margin. Post-tsunami diatom and protozoan assemblages are generally more marine than their pre-tsunami counterparts, suggesting that the study site subsided during earthquakes. Estimates of interseismic and coseismic deformation during one earthquake cycle indicate that less than half of the interseismic uplift was recovered during the earthquake, leading to cumulative, or permanent, uplift of the area.

Key words: Cascadia subduction zone, diatoms, foraminifers, relative sea level, tectonics, tsunamis, British Columbia, late Holocene.

Introduction

The interface between the subducting Juan de Fuca plate and the overriding North America plate at the Cascadia subduction zone (Figure 1a) is currently locked, and the accumulating strain is compressing and deforming the margin of North America (Hyndman and Wang, 1995). The outer section of the continental shelf above the locked zone is being dragged down, whereas the inner continental shelf and outer coast are being uplifted (Hyndman and Wang, 1995). The accumulating strain is likely to be released in the future in a great (M8+) thrust earthquake (Clague, 1997). During such an earthquake the sea floor above the locked zone will be displaced upwards, generating a tsunami, and the coastal zone will suddenly subside, with flooding of low-lying areas. After the earthquake the thrust fault will relock, and the cycle will begin again.

The cycle of elastic deformation at a subduction zone produces land-level changes that can be recognized in the geological record. At estuaries along the Cascadia coast buried marsh soils are abruptly overlain by tidal flat muds, indicating sudden subsidence. The contact between these two facies is commonly marked by a thin layer of sand containing exotic marine microfossils, probably a product of high-energy tsunami waves (Clague and Bobrowsky, 1994; Guilbault *et al.*, 1996; Atwater and Hemphill-Haley, 1997; Benson *et al.*, 1997). Geologists have reconstructed the sequence of late-Holocene plate-boundary earthquakes in Cascadia with some precision using radiocarbon age determinations on plant fossils taken from the top of the buried soils (Table 1).

Although the amount of coseismic subsidence that accompanied the last great earthquake in AD 1700 has been determined at many sites (Peterson *et al.*, 1997), few workers have undertaken an integrated analysis of land-level change over an entire plate-boundary earthquake cycle (Nelson *et al.*, 1996a; 1996b; Shennan *et al.*, 1996; Atwater and Hemphill-Haley, 1997). These workers have
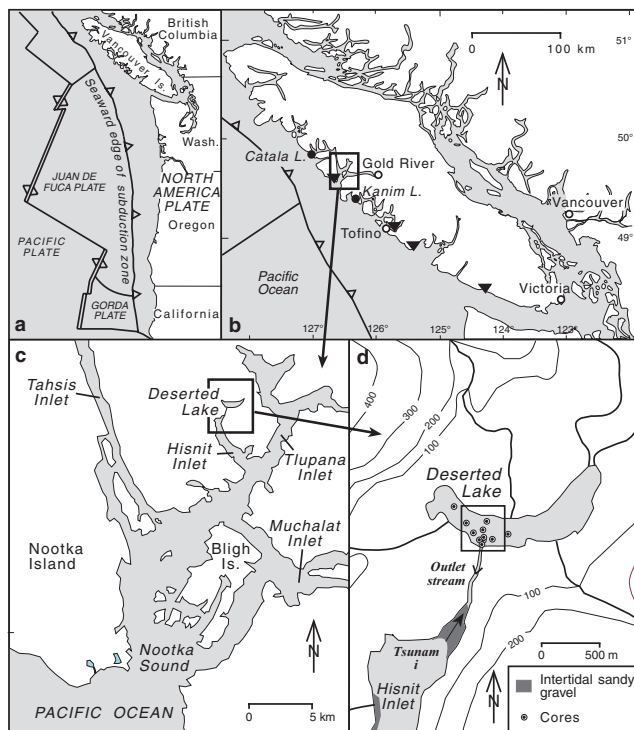
Figure 1 (a) Tectonic setting of the Cascadia subduction zone. (b) Vancouver Island, showing the location of the study area and places mentioned in the text. Triangles show the location of coastal marshes displaying evidence of coseismic subsidence in AD 1700. (c) Nootka Sound, showing the location of Deserted Lake. (d) The environs of Deserted Lake; the rectangle shows the area covered by Figure 2. Elevations are in metres.

Table 1 Ages of plate-boundary earthquakes at the Cascadia subduction zone[1]

Earthquake[2]	Age[3]	
	Best estimate	Range[4]
Y	300	300–300
W	1100	1000–1300
U	1300	1200–1300
S	1600	1300–1700
N	2600	2400–2800
L	3000	2800–3200
J	3400	3400–3500

[1] After Atwater and Hemphill-Haley (1997) and B. Atwater (1997, personal communication).
[2] Plate-boundary earthquakes are designated by names of associated buried soils at marshes on the outer coast of Washington state.
[3] In calendar years before AD 2000.
[4] 95% confidence interval of the calibrated age.

examined sediment sequences underlying intertidal marshes to assess the rate and magnitude of strain buildup and release. In central and southern Cascadia, where most of the studies have been conducted, marshes are extensive and their sedimentary sequences span several thousands of years. In northern Cascadia, however, marshes are of limited extent and have developed largely over the last thousand years (Benson *et al.*, 1997). As a consequence, although the pattern and amount of coseismic subsidence for the last earthquake on the outer coast of Vancouver Island are fairly well known (Clague and Bobrowsky, 1994; Guilbault *et al.*, 1995; 1996), information on the incidence of earthquakes and the pattern of interseismic deformation is very meagre.

Here we examine the lithostratigraphy and biofacies of one core from a low-elevation lake on the west coast of Vancouver Island that has periodically been in contact with the sea. Our aim is to provide a record of deformation spanning several great earthquake cycles in northern Cascadia. We reconstruct changes in the elevation of the lake using diatoms, foraminifers and arcellaceans as proxies of palaeosalinity.

Study site

Our record comes from Deserted Lake (49° 46′N, 126° 30′W), a meromictic lake located in a basin at the head of Hisnit Inlet in the fjord-complex of Nootka Sound on the west coast of Vancouver Island (Figure 1, c and d). Nootka Sound is at the northern end of the Cascadia subduction zone, and Deserted Lake may, therefore, represent a limiting case for detecting tectonic deformation related to the great earthquake cycle.

Deserted Lake covers an area of about 40 ha and is bordered by mountains reaching elevations of more than 1000 m (Figure 1d). Apart from a few open areas where gravel fan-deltas have developed at the mouths of small creeks, the shores of the lake are densely forested. The lake basin has relatively steep sides and a maximum depth of more than 20 m (Figure 2). The pycnocline marking the transition between fresh (<200 μS cm^{-1}) surface waters and saline ($>20\,000$ μS cm^{-1}), anoxic bottom waters lies at about 6 m depth. Gyttja is presently accumulating on the floor of the lake, and little or no sand is being deposited.

The lake is separated from Hisnit Inlet by a bedrock and gravel barrier with a maximum elevation of about 5 m a.s.l. The outlet stream to Hisnit Inlet (Figure 1d) is about 500 m long and is incised 1–2 m into bedrock along part of its upper reach. We used an automatic level to survey the elevation of the lake relative to tidal levels in the sheltered waters of Hisnit Inlet (Figure 1d), which are taken to be equivalent to those at Gold River (Figure 1b) (Fisheries and Oceans Canada, 1997). The elevation of the bedrock outcrop forming the lake sill is 0.97 m above mean sea level, or about 1 m below the upper limit of tides (Fisheries and Oceans Canada, 1997). At present, tidal waters back up the outlet channel about 240 times per year, or about 1.4% of the time (Fisheries and Oceans Canada, 1997). Such incursions of saline water help maintain the meromictic character of the water body.

Methods

Field methods, lithostratigraphy and chronology

A percussion corer (Reasoner, 1993) mounted on a platform constructed from two inflatable boats was used to sample the sedimentary infill of Deserted Lake. Cores were taken from 11 sites radiating from a focus at the lake outlet (Figure 2). The cores were returned to the laboratory for lithological description and sampling. Observations made on the cores include sediment texture, structure, colour and organic constituents, as well as contact depths and characteristics.

Samples of wood, bark, conifer needles and cone bracts were dated by accelerator mass spectrometry (AMS) at Isotrace Laboratory (University of Toronto). This terrestrial plant detritus has been redeposited within the lake and the reported ages therefore represent maximal ages for the enclosing sediment. Radiocarbon ages obtained during this study are cited as calibrated age ranges, determined from the dendro-calibrated data of Stuiver and Becker (1993) using the CALIB 3.0.3A program of Stuiver and Reimer (1993). The longest of the cores (DL-2) was selected for investigation of microfossil assemblages. This core was taken about 80 m from the lake outlet, in 10.5 m of water, on the slope towards the deepest part of the lake (Figure 2).
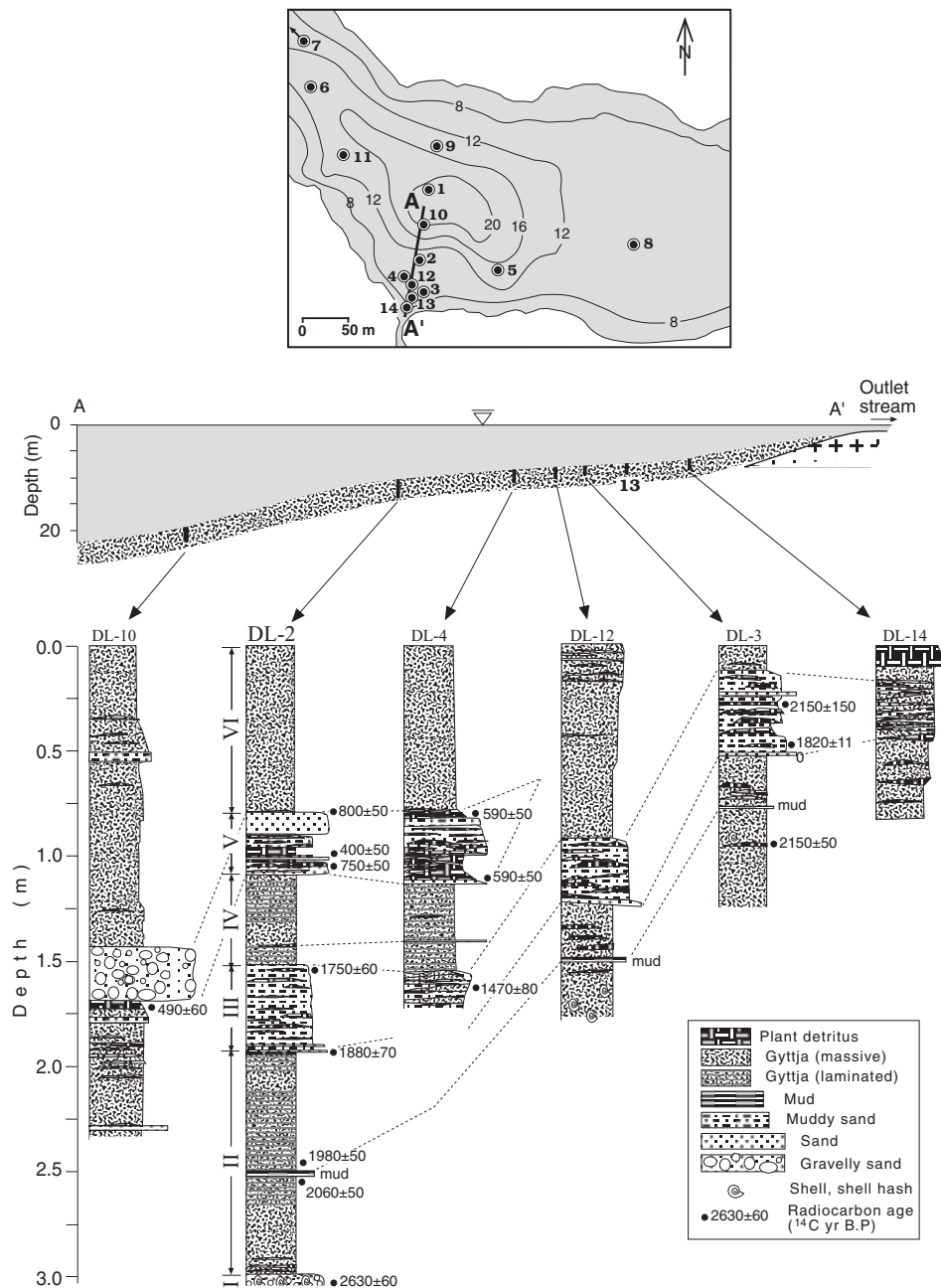
Figure 2 Top: bathymetry of the central basin of Deserted Lake showing core locations, core reference numbers, and position of transect A–A'. Middle: core reference numbers, locations and associated water depths along transect A–A'. Bottom: lithostratigraphy of cores on, or adjacent to, transect A–A'.

Siliceous microfossils (diatoms, chrysophytes and silicoflagellates)

Organic matter in 0.5 g samples from core DL-2 was removed by H_2O_2 digestion in preparation for analysis of siliceous microfossils. The remaining material was dispersed in 250 ml of distilled water, and, after repeated decanting and settling to remove fines and to bring the solution to a near-neutral pH, aliquots of suspended material were dried on glass slides and mounted in Hyrax. At least 300 specimens, or, in diatom-poor materials, the number of valves encountered on 10 random parallel traverses, were identified and counted for each sample.

Diatom identifications were based on descriptions in van der Werff and Huls (1957–74), Hendey (1964), Patrick and Reimer (1966; 1975), Rao and Lewin (1976), Foged (1981), Bérard-Therriault *et al*. (1987), Kling and Håkansson (1988), Laws (1988) and Hemphill-Haley (1993). Species were placed in salinity-tolerance classes following the Halobian system of Kolbe (1927) as modified by Hustedt (1953). The assignments were

based on information derived from the taxonomic sources listed above plus data provided by Haworth (1976), Vos and de Wolf (1988), Pienitz *et al*. (1991) and Nelson and Kashima (1993). Source environments were inferred from the range of salinity tolerance exhibited by the diatom assemblages.

Chrysophyte stomatocysts and skeletal fragments of silicoflagellates encountered during the diatom counts were also enumerated. No attempt was made to discriminate the groups further.

Foraminifers and arcellaceans

Ten cubic centimetre samples were taken from the same depths in the core as those used for diatom analysis. Protozoans were retrieved by washing the samples through a 63 μm mesh sieve. Sieved samples from sandy deposits were dried and the fossils concentrated in a heavy liquid (density = 1.9); aliquots of other samples were obtained with a wet splitter (Scott and Hermelin, 1993). The splits were counted in a wet state using a Wild M5 binocular microscope at a magnification of 50× and were stored

in water with added isopropyl alcohol. A statistically significant number of specimens was counted in as many samples as possible (Patterson and Fishbein, 1987).

The foraminifers, which inhabit marine environments, were grouped into three habitat classes in this study: 'marsh' = species restricted to tidal marshes; 'submarsh' = species living in subtidal and intertidal environments at elevations below the lower limit of the tidal marsh; and 'marsh + submarsh' = species ranging across both of the environments. Arcellaceans were split into two categories: *Centropyxis aculeata*, which is known to tolerate brackish conditions; and other arcellaceans whose saltwater tolerance is minimal or nil.

Results

Lithostratigraphy and chronology

Core DL-2, like most of the other cores recovered from Deserted Lake (Figure 2), consists of massive and laminated gyttja (organic mud and silty organic mud) interbedded with coarser deposits. The boundaries between the strata are commonly sharp.

The lowest cored sediment in DL-2 is gravelly sand and mussel (*Mytilus* sp.) shell hash (unit I). A fragment of wood in unit I (Table 2; TO-6171) provided an age of 2450–2900 yrs BP. This coarse layer is overlain by gyttja (unit II) which contains scattered lenses of plant detritus (twigs, needles) near its base. A thin, sharply demarcated bed of clayey silt occurs within the gyttja at 2.54 m depth. Bracketing ages (Table 2; TO-6169, TO-6170) suggest that this clay layer was deposited about 2000 years ago.

Unit II is sharply overlain by a series of coarse-textured beds (unit III). Two thin sand beds at the base of unit III are separated by a thin layer of sandy gyttja and are overlain by a poorly stratified bed of sandy mud containing woody detritus from 1.89 to 1.53 m depth. Material from the top and base of unit III indicates deposition in the interval 1580–2000 years BP (Table 2; TO-6167, TO-6168).

A second gyttja unit (unit IV), from 1.53 to 1.10 m depth, is overlain by a complex deposit consisting of coarse-textured beds and plant detritus (unit V). The base of this unit consists of five thin (1–5 cm) layers of sand or muddy sand with variable amounts of plant detritus. The clastic beds are separated by thin layers of sandy, woody detritus. The uppermost thin sand layer is overlain

by a thin mud layer which, in turn, is overlain by an 11 cm thick layer of weakly graded, clean fine sand, which is capped by a thin layer of plant detritus. Three samples of reworked forest detritus in unit V yielded ages between 360 and 840 years (Table 2; TO-6164, TO-6165, TO-6166). The uppermost 0.8 m of core DL-2 is massive gyttja (unit VI).

Cores collected closer to the outlet stream than DL-2 contain some, but not all of the above-described units. For example, cores DL-3 and DL-12 contain units II and III, but not unit V, and units IV and VI grade into each other (Figure 2). It appears that part of the upper sequence has been eroded, possibly by a tsunami or storm surge. In contrast, the younger units thicken in deeper water north of DL-2. For example, unit VI, which is 0.8 m thick at DL-2, is almost 1.5 thick in a core taken near the deepest point in the lake (Figure 2: DL-10). At these sites we were unable to core deep enough to penetrate the older units that we sampled at DL-2.

Biostratigraphy: diatoms, chrysophytes and silicoflagellates

A total of 8612 diatoms representing 247 species were enumerated in the 32 samples taken from core DL-2. The basal unit of gravelly sand and shell hash (unit I) contains a rich assemblage of planktonic marine (e.g., *Skeletonema costatum*, *Thalassiosira* spp. (especially *T.* cf. *decipiens*) and *Paralia sulcata*) and benthic marine and marine-brackish (e.g., *Cocconeis scutellum*, *Plagiogramma staurophorum*, *Diploneis smithii*) diatoms (Figure 3). These groups account for more than 80% of the total sample. Silicoflagellate fragments are common, but no chrysophyte cysts were found in the sample.

The sample from the base of the overlying gyttja (2.8 m depth) contains a diatom assemblage similar to that in unit I, but with increased representation of *Skeletonema costatum* and *Cocconeis scutellum* (Figure 3). Marine and marine-brackish diatoms decline to ≤40% of the total count in samples from the middle part of unit II. Benthic species characteristic of brackish environments, such as *Rhopalodia musculus*, *Mastogloia elliptica* and *Navicula peregrina*, and a brackish-fresh planktonic diatom (*Cyclotella* cf. *caspia*) account for most of the remainder (Figure 3). Silicoflagellates are less common in the middle of the unit than at the base. No chrysophyte cysts were encountered in these samples.

The sample from the thin clay layer at 2.54 m depth in unit II contains intact chains of *Skeletonema costatum*, totalling more

Table 2 AMS radiocarbon ages from Deserted Lake

Radiocarbon age (^{14}C yr BP)[3]	Calibrated age range (cal. yr BP)[4]	Core (Figure 2)	Depth (m)[1]	Dated material	Laboratory no.[2]
400 ± 50	360 –570	2	0.98	Cone and bract	TO-6165
490 ± 60	510–610	10	1.73	*Tsuga* needles	TO-6684
590 ± 50	560–700	4	0.48	Twig	TO-6669
590 ± 50	560–700	4	0.78	Twig	TO-6670
750 ± 50	620–790	2	1.04	Twigs	TO-6166
800 ± 50	710–840	2	0.80	Wood fragment	TO-6164
1470 ± 80	1320–1580	4	1.33	*Tsuga* needles	TO-6671
1750 ± 60	1580–1870	2	1.57	Bark	TO-6167
1820 ± 110	1560–2040	3	0.51	*Tsuga* needles	TO-6667
1880 ± 70	1660–2000	2	1.96	Conifer needles	TO-6168
1980 ± 50	1870–2090	2	2.48	Twig	TO-6169
2060 ± 50	1940–2200	2	2.58	Twig	TO-6170
2150 ± 50	1960–2260	3	0.94	Twig	TO-6668
2150 ± 150	1860–2520	3	0.28	*Tsuga* needles	TO-6666
2630 ± 60	2450–2900	2	3.08	Wood fragment	TO-6171

[1] Depth below lake floor.

[2] TO: IsoTrace Laboratory (University of Toronto).

[3] Uncertainties are ± 1σ. All ages corrected to δ^{13}C = –25‰ PDB.

[4] Calibrated age ranges were determined from the dendro-calibrated decadal data of Stuiver and Becker (1993). The ranges represent the 95% confidence interval based on the 1σ error limits of the radiocarbon age (error multiplier = 1).
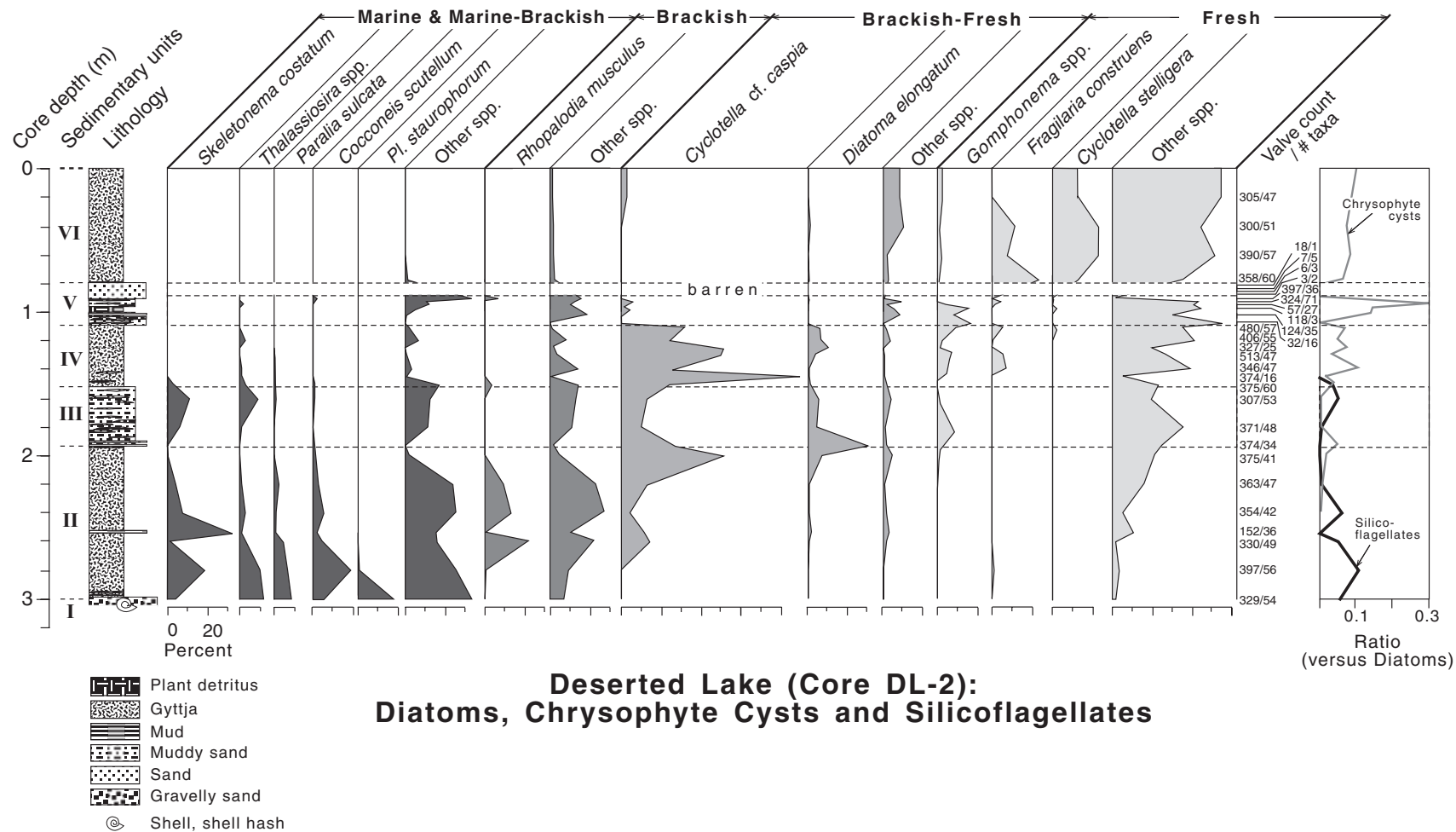
Figure 3 Relative abundance of diatom species, silicoflagellates and chrysophyte cysts in core DL-2. Pl. = *Plagiogramma*.

than 30% of the total diatom count (Figure 3). The sample from the top of unit II contains few (<5%) marine or brackish diatoms. The assemblage at this level is dominated by *Cyclotella* cf. *caspia* and an array of species characteristic of weakly brackish or freshwater habitats (e.g., *Diatoma elongatum*, *Fragilaria ulna*, *Mastogloia pusilla*). Silicoflagellates are absent from this sample and chrysophyte cysts are uncommon.

Samples from unit III contain a diverse diatom assemblage with almost equal amounts of marine and marine-brackish, brackish and fresh-brackish, and fresh diatom species. Saltwater indicator species (e.g., *Thalassiosira* spp., *Paralia sulcata*), increase towards the top of the unit.

The sample taken from the base of unit IV contains a fairly diverse and abundant (>40% total) marine and brackish diatom flora, and relatively abundant silicoflagellates (Figure 3). The marine and brackish components decrease upward, and the siliceous microfossil assemblage in samples from the middle and upper part of unit III resembles that at 2.2 m depth, with common *Cyclotella* cf. *caspia*, brackish-fresh and freshwater diatoms, as well as chrysophyte cysts. Samples from 1.4 and 1.2 m depth, however, show an increase in the relative abundance of diatoms with marine and brackish affinities (e.g., *Gyrosigma balticum*) (Figure 3).

The interbedded muddy sands and organic detrital layers of unit V are dominated by fresh and brackish diatoms, although marine and marine-brackish species increase upward through the unit. Chrysophyte stomatocysts are very common in this unit. The diatom assemblage at the base of the upper clean sand layer in unit V is distinctive. Complete valves of *Navicula digitoradiata*, an epipelic species of brackish-marine intertidal and subtidal environments, are abundant in this sample. The remainder of the samples from this sand contain very few diatoms.

The uppermost unit in the core (unit VI) contains a diverse freshwater diatom assemblage dominated by *Fragilaria construens* (sensu lato) and *Cyclotella stelligera*. Other freshwater indicators such as chrysophyte cysts are fairly common in unit VI, but valves of brackish species are rare and silicoflagellate fragments are absent.

Biostratigraphy: foraminifers and arcellaceans

Foraminifer and arcellacean counts are shown in Figure 4. Unit I and the lowermost sample of unit II are characterized by abundant submarsh forms, mostly *Buccella frigida* and *Trochammina nana*. Arcellaceans are common, the dominant species by far being *Centropyxis aculeata*. The abundant protozoan fauna in the middle part of unit II, from 2.6 m to 2.2 m, is dominated by foraminifers of the marsh + submarsh group, mostly *Ammonia beccarii* and *Miliammina fusca*. Marsh species, in particular *Trochammina inflata*, are relatively common, and arcellaceans, mostly *C. aculeata*, are more frequent.

Relatively few foraminifers and arcellaceans were recovered from the upper part of unit II. The sparse assemblage is dominated by *C. aculeata* and other arcellaceans. The protozoan population in unit III is essentially the same as that in the uppermost sample from unit II, the only differences being the greater abundance of the freshwater indicator *Difflugia oblonga* and the presence of a few foraminifer tests, mostly of the marsh type, in unit III. Unit IV contains large numbers of specimens and is distinguished from unit III by the occurrence at its base of a minor but statistically significant peak of foraminifers, mostly *M. fusca*. The arcellacean content is essentially the same as in the underlying unit. There is a slight decrease in *C. aculeata* and an obvious increase in *D. oblonga* towards the top of unit IV.

Muddy sands and plant detrital layers at the base of unit V contain relatively few foraminifers and arcellaceans (<100 specimens per cm^3). The arcellacean assemblage differs from that in the upper part of unit IV in its lower relative frequency of *D. oblonga*. Very few protozoans were recovered from the overlying

clean sand layer. The assemblage at the base of the clean sand is similar to that in the underlying samples, but in the upper sand the faunal composition gradually becomes similar to that of the overlying unit VI. Arcellacea are abundant in unit VI, with *D. oblonga* slightly more common than *C. aculeata*. No foraminifers were recovered from this unit. This is the only interval in the core in which arcellacean species characteristic of Canadian inland lakes (e.g., Medioli and Scott, 1983; Patterson *et al.*, 1985), such as *Pontigulasia compressa*, *Lagenodifflugia vas*, *Difflugia urceolata*, *D. protaeiformis*, *Cucurbitella tricuspis* and *Lesquereusia spiralis*, are collectively relatively common.

Discussion

Disturbance events

The three coarse, sharply bounded units in core DL-2 (I, III and V) record a series of high-energy pulses of sediment into a low-energy lake or lagoon (Figures 2 and 5). There are also subtle lithological and biostratigraphic changes in DL-2, discussed below, that mark three minor perturbations.

The basal sediment in the core (unit I) is gravelly sand containing shell (particularly mussel) hash and plant detritus. This layer is capped by a thin mud layer containing abundant (30–50%) coarse plant detritus. The diatoms and the protozoans in unit I are strongly marine in their affinities. On the basis of this evidence we infer that unit I was deposited during a short-lived inundation of the Deserted Lake basin by a landward-directed wave from an offshore source that scoured gravel, sand and shell from beaches and rocks, and eroded plant detritus from the forest floor on either side of the outlet stream. The location of core DL-2, on a slope rising towards the outlet (Figure 2), precludes deposition by a stream flowing into the lake from the east, north or west. A piece of wood from unit I gave a calibrated age of 2450–2900 yrs BP, which corresponds to the estimated age of Cascadia plate-boundary earthquake N (Table 1). We conclude that unit I is the product of the tsunami generated by this event.

The basal sample from the poorly stratified sequence of sand and plant detritus of unit III has diatom and protozoan assemblages that are akin to those in the underlying gyttja; only 2% of the diatoms, for example, are species characteristic of marine or brackish environments. The source area for this material was probably the lake bed upslope from DL-2. Diatoms with marine and brackish affinities increase upward through the unit, constituting about 50% of the sample in the uppermost horizon (Figure 5). This material was probably derived partly from the lake bed and partly from shallow-water deposits in Hisnit Inlet which were carried into the lake by a landward-directed wave. We did not find planktonic foraminifera in this unit, or elsewhere in core DL-2, but in a nearshore setting such as Hisnit Inlet they must be very rare. A bark fragment from the top of unit III yielded an age of 1580–1870 years BP, which is a maximum for the age of the deposit. Unit III may be the product of the tsunami generated by the next Cascadia earthquake in the sequence (event S), which occurred about 1600 years ago.

The low numbers of foraminifers and arcellaceans at the base of unit V suggests that the sediment was deposited rapidly. The diatom and protozoan assemblages at the base of unit V are similar to those in the underlying gyttja, but diatoms with marine and marine-brackish affinities again increase towards the top of the unit.

The thick clean sand layer at the top of unit V, which is capped by a layer of plant detritus, has a protozoan content similar to that of the underlying material. The abundance of *Navicula digitoradiata* at the base of this layer is noteworthy, as no other specimens of this diatom were encountered in the other samples from core DL-2. The middle part of the sand is essentially barren. In
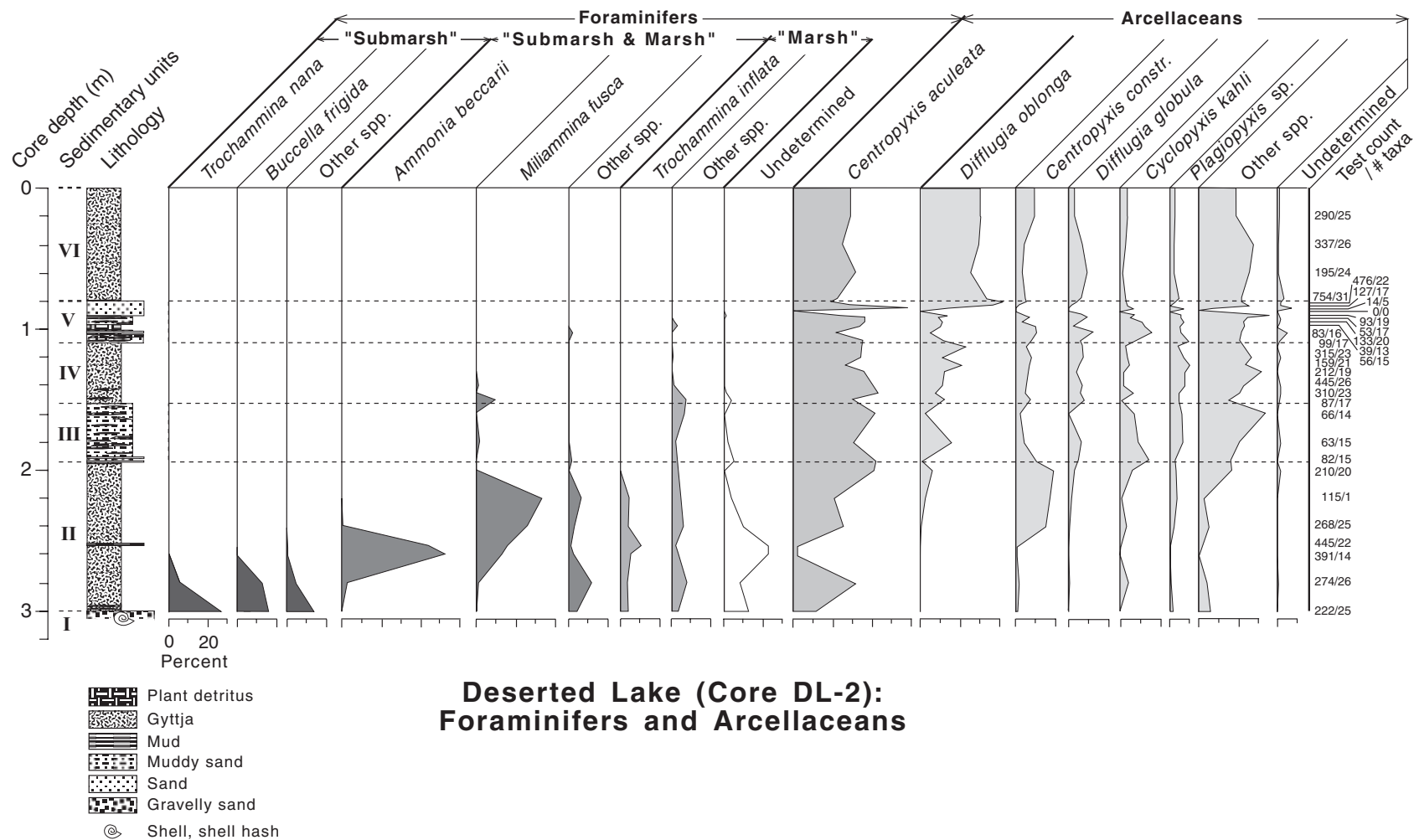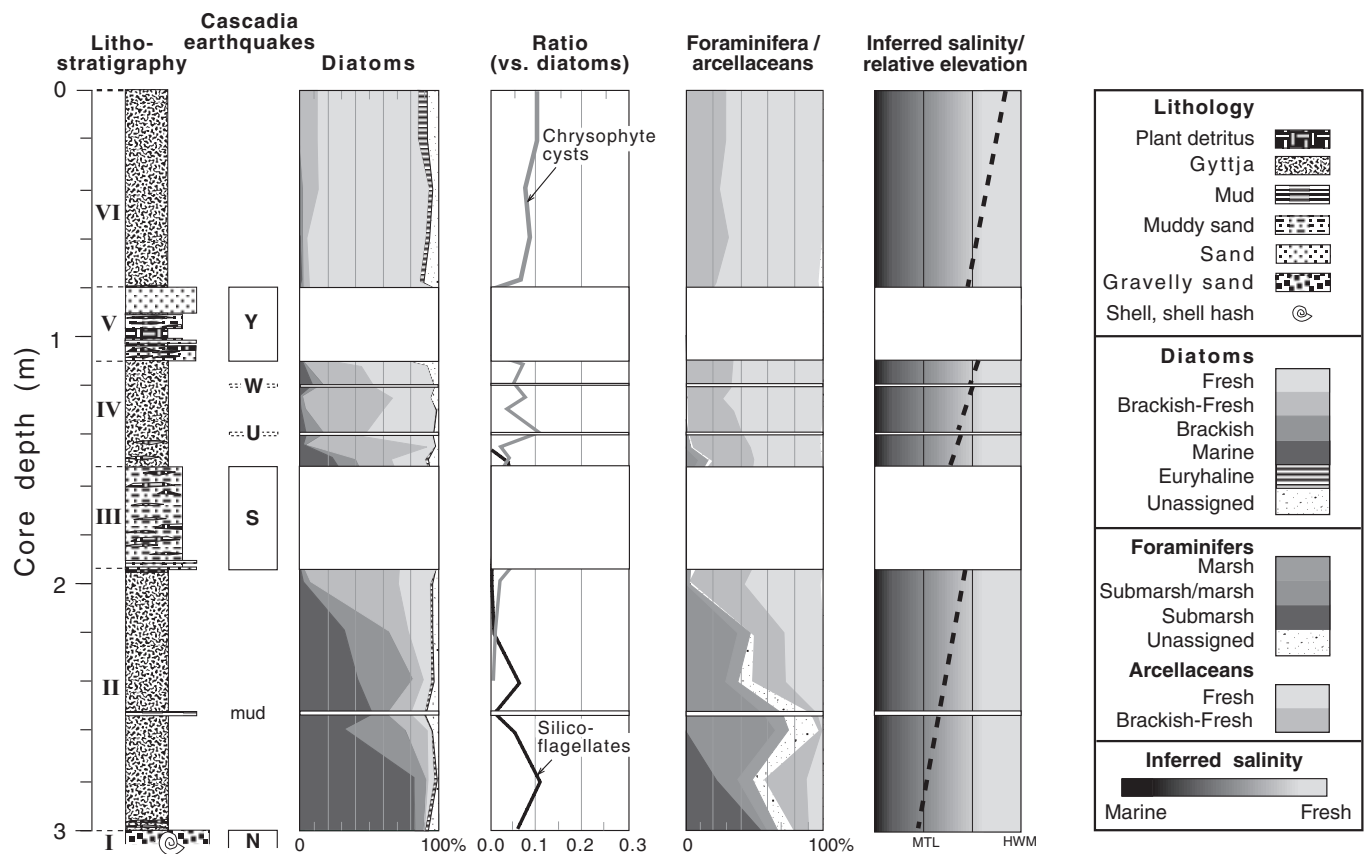
Figure 4 Relative abundance of foraminifers and arcellaceans in core DL-2.

Figure 5 Summary of lithostratigraphic and biostratigraphic evidence from core DL-2, showing inferred changes in salinity and approximate elevation (dashed line) of Deserted Lake during interseismic periods. MTL = mean tide level; HWM = high water mark.

contrast, samples from the upper part of the bed have an arcella-cean composition similar to that of the overlying gyttja, perhaps as a result of burrowing by these infaunal species.

The youngest of the three age determinations on reworked material from unit V gives a maximum age for the deposit of approximately 360–570 yrs BP. We interpret unit V to be the product of the tsunami generated by the Cascadia earthquake of AD 1700. Each of the thin sandy interbeds at the base of this unit was probably deposited by a single wave in the tsunami train. As with unit III, the microfossil content of the basal sediment indicates that it was probably scoured from the stream channel and lake bed upslope of DL-2, but the increased admixture of marine diatoms towards the top suggests that later waves were also introducing material into the lake from intertidal areas at the head of Hisnit Inlet. In contrast, the weakly graded, thick layer of clean sand at the top of unit V seems to be derived entirely from an offshore source area and to have been deposited by a single land-ward-directed surge of water. We interpret the thin mud layer between these two coarse phases of unit V to represent a short period of quiescence between disturbance events. If the bracketing coarse layers are indeed tsunami deposits, their contrasting litho-logies and diatom assemblages suggest that the waves that pro-duced them had different characteristics.

The oldest of the three minor disturbances is marked by the clayey silt layer at 2.54 m depth. It strongly resembles a deposit of about the same age at Kanim Lake (Hutchinson *et al.*, 1997), some 35 km to the south. We consider it likely that both deposits are the products of severe floods or debris flows, although the presence of abundant intact chains of the marine diatom *Skele-tonema costatum* in the clayey silt layer in Deserted Lake is some-what puzzling in a deposit with an inferred terrestrial origin. This species dominates the phytoplankton in the coastal waters of Bri-tish Columbia during late winter and spring (Harrison *et al.*, 1983), and its presence in the clayey silt layer possibly marks the

die-off of a *Skeletonema* bloom in the Deserted Lake embayment caused by high turbidity in the water column following a severe flood or debris flow.

The evidence for other minor disturbance events is more tenu-ous. There are spikes in the abundance of marine, marine-brackish and brackish diatoms at 1.20 and 1.40 m depth (Figure 5). The latter is associated with a layer of forest detritus and may be cor-relative with a very thin sand layer at the same depth in DL-4. These events are undated, but if they are tsunami deposits they may be the products of Cascadia plate-boundary earthquakes W and U, which date to about 1100 and 1300 years ago (Table 1).

Permanent deformation

The microfossil assemblages in core DL-2 record a change in the character of Deserted Lake in the last 3000 years from a tidal lagoon that was strongly influenced by the ocean to the present-day meromictic lake with its fresh epilimnion. The tidal range has almost certainly been constant in the late Holocene on this open fjord coast, so this marine to freshwater transition must record a relative sea-level fall of 1–2 m. This change may be the product of regional eustatic factors or land-level fluctuations. The latter may be associated either with postglacial isostatic recovery or with tectonic deformation of the convergent margin.

If the changeover from a marine to a freshwater environment at Deserted Lake over the last few thousand years is a result of eustatic sea-level fall, then sites in the Pacific Northwest beyond the isostatic influence of the Cordilleran ice sheet, such as the Oregon coast, or sites located landward of the convergent margin, such as the Fraser Lowland on the mainland coast of British Columbia, would also have been subject to a fall in sea level. Geological evidence, however, shows that relative sea level in these areas has risen by about 1 m (at the Fraser River delta, just south of Vancouver) to 2–3 m (on the central Oregon coast) in the last three millennia (Hutchinson, 2000).

Isostatic rebound can also be ruled out as an explanation for relative sea-level fall in the late Holocene. Output from a regional Earth model constrained by Cordilleran ice-sheet history, lithospheric thickness and upper mantle rheology shows that glacio-isostatic relaxation of the west coast of Vancouver Island was largely complete by 11 000 ^{14}C yrs BP (T. James (Geological Survey of Canada), personal communication, 1999). Model output indicates that present-day isostatic uplift rates in the Nootka Sound area are negligible (<0.02 mm a^{-1}; James *et al.*, 2000). Three thousand years of uplift at this rate would generate only 0.06 m of sea-level fall.

The final possibility is that regional sea-level fall results from tectonic activity at the Cascadia margin. Although elastic rebound theory demands that the strain accumulated during interseismic periods is recovered by coseismic recoil, in reality there is always residual, unrecovered strain following plate-boundary earthquakes (Thatcher, 1984a; 1984b). Those parts of the overriding plate which are subject to interseismic uplift and coseismic subsidence are thus gradually uplifted over the course of several earthquake cycles, leading to permanent deformation of convergent margins (Thatcher, 1984a; 1984b).

The fact that the inferred tsunami deposits in core DL-2 are bracketed by microfossil assemblages which record coeval subsidence (see below) provides strong support for the hypothesis that the overall fall in sea level revealed in DL-2 is a result of residual tectonic deformation of this coastal margin. If we assume that the regional eustatic sea level has risen (as has been the case at the Fraser River delta) by about 1 m in the last 3000 years, the 1–2 m fall in relative sea level recorded at DL-2 must be the cumulative product of 2–3 m of uplift in this interval; an average annual uplift rate of 1 to 1.5 mm a^{-1}.

Interseismic and coseismic deformation

Microfossil assemblages from gyttja units provide evidence of relative sea-level change between disturbance events. The biofacies from the lower part of unit II is strongly marine in character, although, as in the inner reaches of Effingham Inlet (Patterson *et al.*, 2000), arcellaceans are relatively common. The arcellaceans are probably reworked; this is not unexpected in a small basin surrounded by mountains with a maritime climate. More than 80% of the diatoms in the basal gyttja sample belong to species tolerant of high salinities, marine silicoflagellates are abundant, and three of the major foraminifer species in this sample are common in shallow subtidal environments along the British Columbia coast (Reinhardt *et al.*, 1996; Patterson *et al.*, 2000). We conclude that about 2600 years ago Deserted Lake was a marine embayment with a threshold below mean sea level. Strong tidal flushing is to be expected in such a setting. The DL-2 core site must have been subtidal at this time, with water depths at the site ranging from 8 to 12 m, depending on tidal state.

Samples from the middle of unit II are dominated by marine-brackish, brackish (e.g., *Rhopalodia musculus*) and weakly brackish (e.g., *Cyclotella* cf. *caspia*) diatoms, 'submarsh + marsh' (e.g., *Ammonia beccarii* and *Miliammina fusca*) and 'marsh' foraminifers, and *Centropyxis aculeata*. Although the presence of typical marsh foraminifers in a subtidal setting seems somewhat paradoxical, a search of Murray (1991) and of references therein shows that typical marsh species, particularly *Trochammina inflata*, can live subtidally. The relative abundance of *Cyclotella* cf. *caspia* indicates that surface waters could on occasion be oligosaline. The entire assemblage indicates that tidal inflow into this lagoon was more restricted 2000 years ago than it had been 600 years earlier. We conclude that the Deserted Lake sill was relatively higher, and was probably located in the mid- to upper intertidal zone at that time.

In contrast, about 90% of the diatoms in the sample from the top of unit II are species characteristic of weakly brackish or freshwater environments (Figure 5). In addition, stomatocysts of freshwater chrysophytes and arcellaceans are common in this sample, and there are few foraminifers and no silicoflagellate fragments. This biofacies records a lacustrine environment subject to infrequent tidal incursions. The elevation of the lake sill at the time of deposition (about 1700 years ago) was probably a little below the present sill. It should be noted, however, that the uppermost gyttja of unit II may have been lost to erosion by the ensuing tsunami, so this elevation estimate represents a minimum height for the sill prior to earthquake S.

Unit II spans a period of about 1000 years. Our estimates of relative sill elevation and eustatic sea-level rise suggest that there was about 2 m of crustal uplift in this interval, equivalent to an average annual uplift rate of about 2 mm a^{-1}. The relatively long-lived brackish phase in the middle of this interval suggests that uplift during this period was more or less uniform, and was not concentrated in the period immediately following earthquake N (Figure 5).

The relative abundance of marine and brackish diatoms, silicoflagellates and marsh foraminifers in the samples from the base of unit IV implies that the surface waters of Deserted Lake were fairly saline immediately after the inferred tsunami associated with earthquake S. Although tsunami inundation would probably disturb the meromictic character of Deserted Lake, this effect would be very short-lived in the perhumid maritime climate of Vancouver Island, and we conclude that saltwater could enter the palaeolagoon with relative ease in the aftermath of the tsunami. A similar post-tsunami increase in marine indicator species was found at Catala Lake (Figure 1b) (Clague *et al.*, 1999). Catala Lake is separated from the sea by a gravel barrier, and the increase in salinity may have been a product of barrier erosion by the tsunami. Similarly, the freshening tendency recorded at Catala Lake between tsunamis may be primarily a result of barrier reconstruction during the interseismic interval. At Deserted Lake, however, the lake sill is bedrock-controlled, and the post-tsunami increase in saltwater intrusion must have resulted from regional coseismic subsidence. The diatom and protozoan biofacies at the base of unit III suggest that at that time the sill lay several decimetres below its present elevation. Given that prior to earthquake S the sill was close to its modern position, the area may have subsided by as much as 0.5–1 m during the earthquake. Subsidence during the earthquake would, of course, allow subsequent tsunamis to inundate the lake basin more readily.

The diatom and protozoan biofacies of unit IV indicate a gradual freshening-up unit. In the uppermost sample >95% of the diatoms are fresh and fresh-brackish species. In the case of the protozoans the freshwater arcellacean *Difflugia oblonga* becomes more abundant upward and *Centropyxis aculeata* less so.

The approximately 1300-year interval between earthquake S (about 1600 years ago) and earthquake Y (300 years ago) is represented by 0.40 m of gyttja (unit IV), equivalent to an average sediment accumulation rate of 0.3 mm a^{-1}. In the most recent interseismic period (since AD 1700 = unit VI), it has averaged 2.7 mm a^{-1}. Although unit IV has undoubtedly compacted as a result of dewatering, and the rate of sediment accumulation in Deserted Lake has undoubtedly fluctuated during the Holocene as a result of climate change, the low apparent rate of sedimentation between earthquakes S and Y implies that this section of the core was truncated by the tsunami of AD 1700. The elevation of the lake sill immediately before the AD 1700 earthquake is therefore unknown, but was probably at least as high as at present.

The elevation of the lake just after the AD 1700 event is constrained by the low incidence of marine and brackish indicator species at the base of unit VI. It is clear that Deserted Lake was not lowered into the mid-tidal zone, but the presence of a few brackish diatoms suggests that it may have been slightly lower than at present. Guilbault *et al.* (1995; 1996) estimate that about

0.5–0.7 m of subsidence occurred during this event at Tofino, some 80 km to the southeast (Figure 1b). The amount of subsidence at Deserted Lake appears to have been substantially less than this.

The upper part of unit VI is the freshest part of the core. Diatoms in this unit are almost entirely freshwater species, the arcellacean *Difflugia oblonga* is generally dominant over *C. aculeata*, and arcellaceans characteristic of inland lakes are present in low numbers. The high density of protozoan specimens and the rather uniform microfossil assemblages suggest an environment that is stable and largely insulated from the ocean. However, the continuing abundance of *C. aculeata*, low aggregate percentage of 'inland' arcellaceans, and the persistent presence of a few brackish diatoms implies that tidal inflows still entered the lake throughout the 300 years during which unit VI was deposited.

Conclusions

Three high-energy depositional events are recorded in core DL-2 by coarse or medium-textured sediments (gravel to muddy sand), which are interbedded with, and commonly overlain by, layers of plant detritus. These sediments closely resemble those that we have identified as tsunami deposits in other lakes on the west coast of Vancouver Island (Hutchinson *et al.*, 1997; Clague *et al.*, 1999). Based on the similarities between the ages of these sediments and the ages of inferred great earthquakes at the Cascadia subduction zone (Table 1), we conclude that the deposits are the product of tsunamis generated by plate-boundary earthquakes. We tentatively correlate unit I with earthquake N, which occurred about 2600 years ago (Table 1). Unit III may correspond to event S (about 1600 years ago), and the basal part of unit V is tentatively assigned to earthquake Y in AD 1700.

Small-scale changes in relative sea level resulting from tectonic deformation can be detected from variations in the abundance of taxa that are sensitive to changes in lake salinity in the muddy gyttja units in core DL-2. It is clear from the diminishing abundance of marine indicators (marine and brackish diatoms, silicoflagellates and foraminifers) directly above each of the inferred tsunami deposits (Figure 5) that the strain buildup in the interseismic periods is not recovered fully after major earthquakes, and that the Deserted Lake area is gradually rising. We estimate the average rate of permanent uplift for the late Holocene to be 1–1.5 mm a^{-1}, equivalent to the value cited for the west central coast of Vancouver Island by Clague *et al.* (1982) and Friele and Hutchinson (1993).

Gyttja deposited before the AD 1700 earthquake appears to have been truncated by the ensuing tsunami; consequently, the amount of deformation associated with this event cannot be determined. Changes in microfossil assemblages allow a fairly precise estimate of tectonic deformation only in the case of earthquake S. From these variations we estimate that some 2 m of uplift occurred in the 1000-year interval prior to this great earthquake, and that this coastal area appears to have dropped about 0.5–1 m during the event. This implies that less than half of the interseismic deformation is recovered by coseismic subsidence in this area. It should be noted, however, that because Deserted Lake is near the northern margin of the Juan de Fuca plate the elevation changes resulting from coseismic subsidence here are probably smaller than those along the central segment of the subduction zone.

We consider the approach outlined in this paper to be of considerable utility in extending the historical record of interseismic and permanent deformation in other coastal settings on convergent margins, particularly in areas where intertidal marshes have a limited timespan.

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