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Reconstructing the Tsunami Record on an Emerging Coast: a Case Study of Kanim Lake, Vancouver Island, British Columbia, Canada

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ABSTRACT

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A pilot study was conducted at Kanim Lake on the emerging coast of western Vancouver Island, British Columbia, to assess the efficacy of using lake sediments to determine tsunami run-up and recurrence. Sediment sequences in lakes near the coast can complement tsunami records derived from deposits underlying intertidal marshes. Marshes on emerging coasts are uncommon, of limited areal extent, and, most importantly, their deposits have a short lifespan. Tsunami deposits in lakes are less susceptible to bioturbation and erosion and, generally, can be more accurately dated than similar deposits in marshes and other terrestrial settings. An inferred tsunami deposit in Kanim Lake has distinctive lithological characteristics and contains marine and brackish-water microfossils. Kanim Lake also illustrates some of the limitations in using lakes to reconstruct tsunami run-up and recurrence. Although the lake has been in the potential run-up zone for tsunamis triggered by great earthquakes on the nearby Cascadia subduction zone for the last 3,500–4,000 years, it apparently has been inundated by only one tsunami in this period. This event probably occurred about 2,800 years ago. Tsunamis since that time have failed to reach Kanim Lake as the lake basin has continued to rise through the run-up zone and the distance to the sea has increased. The development of dense forest stands on the progressively widening reach between the sea and the lake has probably been the most important factor in limiting tsunami access to the site.

ADDITIONAL INDEX WORDS: *Cascadia subduction zone, diatoms, lakes, relative sea level, tsunami deposits, Vancouver Island.*

INTRODUCTION

There have been no great (moment magnitude $M_w \geq 8$) earthquakes on the Cascadia subduction zone (the interface of the Juan de Fuca and North American plates [Figure 1a]) in the historic period, but a variety of geological and geophysical data suggest that several such events have occurred over the course of the late Holocene (ATWATER *et al.,* 1995; DARIENZO and PETERSON, 1995; HYNDMAN, 1995; ATWATER and HEMPHILL-HALEY, 1996). The most convincing evidence in support of this conclusion is derived from studies of stratigraphic sequences in estuaries from northern California to southern Vancouver Island. Deposits in the high intertidal zone of these estuaries display repetitive sequences of peats abruptly overlain by intertidal muds. The peat layers accumulated slowly in intertidal marshes and floodplain forests, whereas the overlying muds are inferred to have been deposited in lower intertidal environments following sudden coseismic subsidence. In many cases the contacts between the buried marsh peats and the overlying muds are marked by thin, landward-fining sand layers that are inferred to have been deposited by tsunamis generated by these earthquakes (ATWATER, 1987, 1992; DARIENZO and PETERSON, 1990; CLAGUE and BOBROWSKY, 1994b).

Estimates of the height of tsunami waves generated by a great earthquake along the British Columbia segment of the Cascadia subduction zone were made by NG *et al.* (1990) and WHITMORE (1993) from numerical models. They concluded that 4–5 m high waves would reach the outer coast of Vancouver Island 10–15 minutes after the earthquake, and that these would amplify to heights of about 15 m at the heads of some west coast fjords. In the sheltered waters in the lee of Vancouver Island the height of the tsunami waves would progressively diminish, with waves ≤21 m reaching Victoria and Vancouver (Figure 1b) some 3–4 hours after the earthquake.

Although these results convey the extreme nature of the hazard and the short emergency response times for communities on the British Columbia coast, it should be noted that numerical models but may not provide reliable forecasts of

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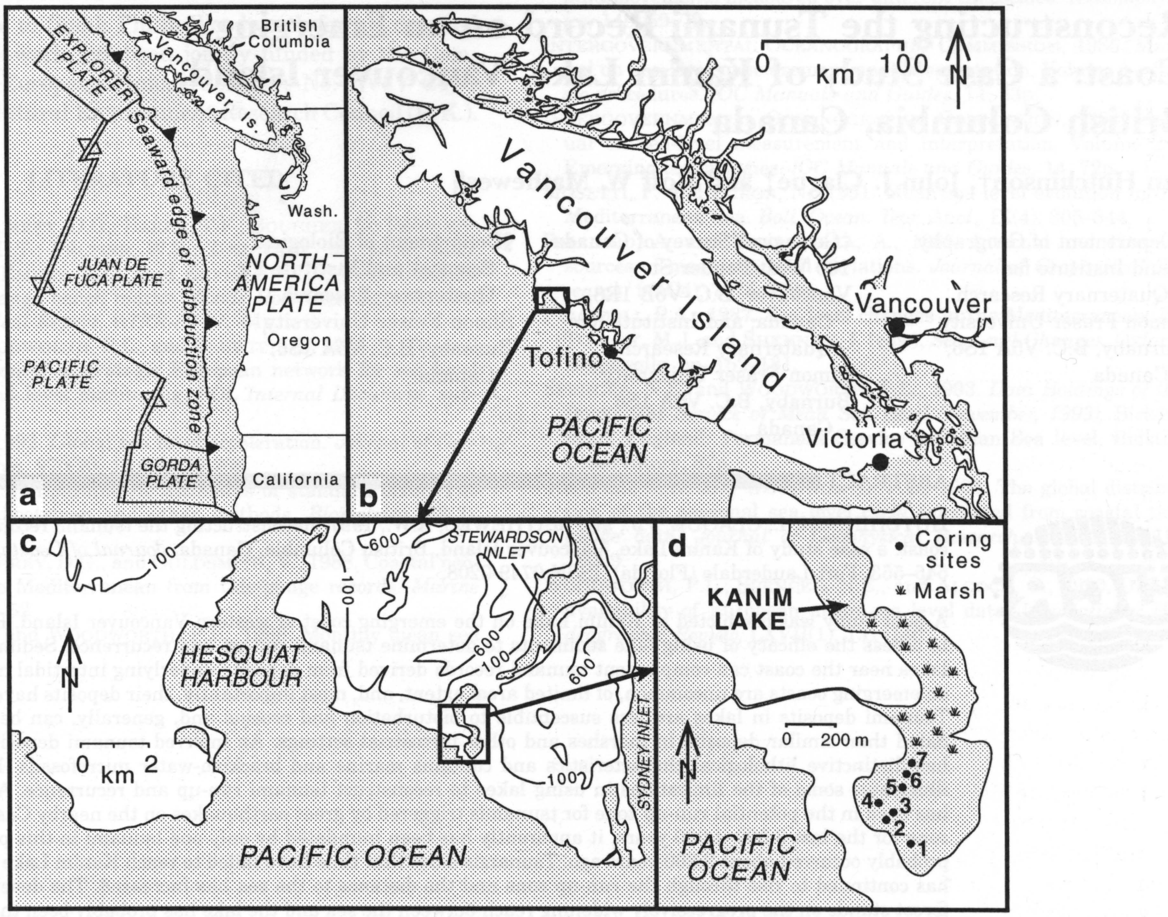

Figure 1. The study area. (a) Tectonic setting of the Cascadia subduction zone. (b) Vancouver Island, showing the location of the study site and places mentioned in the text. (c) The environs of Kanim Lake. Elevations in metres. (d) The southern basin of Kanim Lake showing core sites.

tsunami magnitude. KANAMORI and KIKUCHI (1993) point out that tsunamis display a wide range of individual behaviour and that the magnitude of a subduction zone earthquake is a relatively poor predictor of sea-surface displacement. Relatively small tsunamis may result from large thrust earthquakes if the fault rupture occurs instantaneously and at considerable depth on a non-accretionary margin (*e.g.* Japan, October 1994). Some moderately large tsunamis (*e.g.* Nicaragua 1992) are thought to arise from slow rupture of the plate interface at relatively shallow depths. Extremely large tsunamis (*e.g.* Sanriku 1896) can result if fault blocks in the accretionary prism slump during the earthquake (KANAMORI and KIKUCHI, 1993). Given the variety of modes of interface deformation and sea-bed displacement, and the complex effects of nearshore shoaling and coastal refraction, simulation models may only provide an approximation of the tsunami hazard at specific coastal locations. More accurate estimates of the magnitude and frequency of tsunamis in a coastal area may be obtained from the stratigraphic record.

This paper examines the feasibility of documenting tsunami run-up and recurrence on an emerging coast from the depositional records of nearshore lakes. The paper includes the results of a reconnaissance study of Kanim Lake, located on the central west coast of Vancouver Island at the north end of the Cascadia subduction zone (Figure 1).

RECONSTRUCTING TSUNAMI OCCURRENCE FROM THE GEOLOGICAL RECORD

Previous geological investigations of tsunami run-up and recurrence have largely focussed on the analysis of potential tsunami deposits in modern or relict intertidal marshes and lagoons (*e.g.* LONG *et al.*, 1989; MINOURA and NAKAYA, 1991; ATWATER and MOORE, 1992). The characteristic deposits in these environments are slackwater muds and peats. These low-energy deposits may be interbedded with laterally continuous, coarser material laid down during floods, windstorms, storm surges, or tsunamis. Fluvial deposits commonly coarsen and thicken landward and usually contain a sparse freshwater microfossil assemblage (*e.g.* HUTCHINSON *et al.*, 1995). Strong onshore winds can transport sand from the upper shoreface and dunes and redeposit it in back-barrier settings, but such eolian deposits are likely to be barren of microfossils. Storm surge and tsunami deposits consist of landward-thinning and -fining sheets of sand and, rarely, gravel containing brackish and marine microfossils (*e.g.* REINHART and BOURGEOIS, 1987; DARIENZO and PETERSON, 1990; DAWSON *et al.*, 1991; ATWATER, 1992; DARIENZO *et al.*, 1994; HEMPHILL-HALEY, 1995). These deposits may be difficult to distinguish.

The Pacific coast of southern Washington and northern Oregon, where much of the geological work on great Cascadia earthquakes has been done, is indented by large estuaries which are partly enclosed by sandy barrier beaches and spits. The barrier beach and spit complexes provide an abundant supply of sand for entrainment by tsunamis, and the extensive intertidal marshes of the upper estuaries provide a suitable environment for sand deposition. Rising sea level along the Washington and Oregon coast during the Holocene (HUTCHINSON, 1992) has led to continuous estuarine infilling, and burial and preservation of tsunami deposits.

In contrast to that to the south, the coastal plain along the northern sector of the Cascadia subduction zone (Vancouver Island) is narrow. Mountain ranges backing the coastal plain are indented by narrow, steep-sided fjords. Patches of intertidal marsh are restricted to small areas at the heads of sheltered inlets or the foreshores of fjord-head deltas. The latter are high-energy fluvial systems; tsunami deposits in these environments are rapidly destroyed through channel migration and avulsion. Moreover, unlike areas further south, relative sea level has fallen on the west coast of Vancouver Island during late Holocene time (CLAGUE *et al.*, 1982; FRIELE and HUTCHINSON, 1993). This area is rising relative to the sea (taking into account eustatic sea-level rise) at a rate of about 1 m ka^{-1}. Intertidal marshes on this coast, which occupy a vertical range of about 1 m, therefore have a lifespan of about 1,000 years before they emerge from the intertidal zone (Figure 2). Present-day marshes are therefore likely to record only those tsunamis that have occurred in the last millenium. On Vancouver Island uplifted marshes are rapidly colonized by forest, and their buried tsunami deposits are obscured or destroyed by bioturbation and erosion. For these reasons, there are only deposits from three tsunamis in the marshes near Tofino (CLAGUE and BOBROWSKY, 1994a, b) (Figure 1b). CLAGUE and BOBROWSKY (1994a, b) attributed these deposits to the Alaska tsunami of 1964, a 100–400 year old Cascadia tsunami, and a tsunami from an unknown source sometime between 500 and 800 years ago.

To document older events on emerging coasts like western Vancouver Island, depositional sites above the limit of tides must be investigated. The probability of inundation in the run-up zone can be illustrated by a simple graphical model based on average tidal conditions (Figure 2). If, for instance, a 5 m wave (the maximum suggested by NG *et al.* (1990) for a local tsunami on the west coast of Vancouver Island) coincided with low tide (-2 m asl), only sites below $-2 + 5 = 3$ m asl in the potential run-up zone would be inundated. Sites below this elevation thus would be inundated at all states of the tide (inundation probability (p) = 1; Figure 2). If this model wave coincided with high tide (2 m asl), all sites below $2 + 5 = 7$ m asl (p = 0; Figure 2) would be within the potential run-up zone. The probability of inundation between

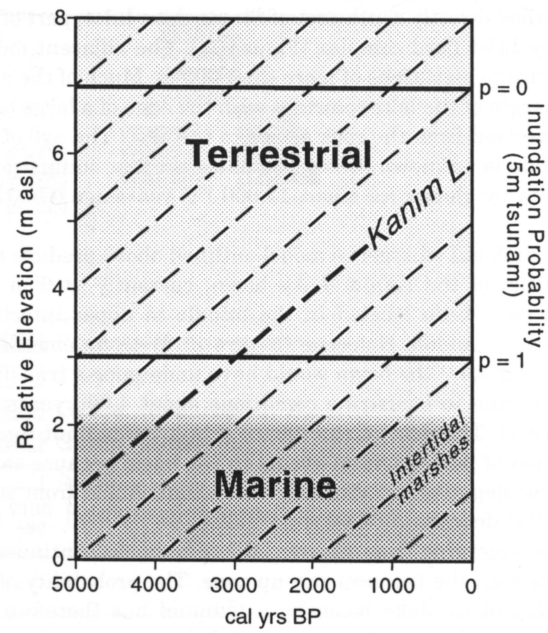

Figure 2. Graphical model of inundation probability as a function of site elevation in the run-up zone for a 5 m tsunami. Hatched lines indicate the relative elevation of sites with respect to sea level for the last 5,000 years, assuming an uplift rate of 1 m ka^{-1}. The study site (Kanim Lake), currently at ± 6 m asl, emerged from the intertidal zone (upper limit ± 2 m asl in sheltered locations), about 4,000 years ago.

these limits (3 m and 7 m asl) decreases with increasing elevation.

On the coastal plain of western Vancouver Island the supratidal run-up zone is occupied by dense temperate rain forest and scattered bogs and lakes. Tsunamis that penetrate the forest may leave patchy deposits, but these will be rapidly obliterated by treefall and erosion. Lakes, in particular those with low fluvial inputs, appear to be the best sites for paleotsunami investigations. Not only is the lithology of a tsunami deposit likely to be distinct from that of low-energy, autochthonous lake deposits such as mud and gyttja, but the associated marine or brackish-marine microfossils will contrast sharply with the freshwater forms native to the lake. Moreover, a tsunami deposit has a much higher probability of burial and preservation in a lake than in the neighbouring forest.

STUDY SITE

Kanim Lake (49° 24′ N, 126° 20′ W; Figure 1c) lies in a partially bedrock-rimmed basin at 6 m asl. The southern part of the lake (hereafter referred to as the southern basin) is shallow (<2 m water depth). The southern margin of this basin is separated from the Pacific Ocean by a rocky isthmus 200 m wide with a minimum elevation of 10 m asl. The central part of the lake is occupied by patches of aquatic vegetation and emergent marsh (Figure 1d); water depths in this area are <1 m. A 1 km long outlet stream flows to the west along a narrow channel from the central part of the lake to the sea (Figure 1d). The larger northern basin occupies a narrow valley directly southwest of Stewardson Inlet, part of the Sydney Inlet fjord complex (Figure 1c). The adjacent mountains rise to elevations of more than 600 m. Much of the eastern margin of the lake coincides with the edge of a large landslide derived from the rock slopes to the east. The age of the landslide is unknown, but it postdates the last occupation of the area by glacier ice about 13,000 ^{14}C years ago (CLAGUE, 1981).

The regional emergence model outlined above predicts that the outlet of the Kanim Lake basin, currently at 6 m asl, would have been located in the middle to upper intertidal zone in the middle Holocene (Figure 2). With an emergence rate of 1 m ka^{-1} the basin would have undergone a transition from marine to lacustrine conditions about 4,000 years ago (Figure 2). The date of emergence sets a lower limit for the detection of paleotsunami events in the basin because sandy tsunami deposits are virtually indistinguishable from some intertidal deposits.

After emerging from the sea, the lake basin continued to rise through the tsunami run-up zone. The probability of inundation of the lake basin by a tsunami has therefore declined through time from the date of emergence to the present. During its early history, Kanim Lake would likely have been inundated by any large tsunami striking this coast. During the last 1000–2000 years, however, it may have been too high to be inundated by a tsunami, unless the maximum wave coincided with high tide.

METHODS

To reconstruct the paleotsunami record from Kanim Lake, undisturbed cores were taken in the southern lake basin and near the lake outlet (Figure 1d) with a Livingstone corer operated from a platform constructed from two inflatable rafts. This lightweight coring equipment was necessitated by the remoteness of the site, which can be accessed only by helicopter. The cores were described in the field; critical material for diatom analysis and radiocarbon dating was returned to the laboratory.

In preparation for diatom analysis, organic matter in 1 g samples was removed by H_2O_2 digestion and the remaining material dispersed in 250 ml of distilled water. 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. The diatom assemblage of each sample was determined at $\times$1,000 by counting the first 200–300 specimens encountered in random parallel traverses. Taxonomic 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), 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). These 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 determined from the range of salinity tolerance exhibited by the diatom assemblages.

Radiocarbon ages obtained during this study, or quoted from previous work, are cited in radiocarbon years before AD 1950 (^{14}C yr BP $\pm$ 1σ). Marine shell ages were normalized to $\delta^{13}C = -25‰$ PDB by the laboratory, and a further regional oceanic reservoir correction of 390 years (STUIVER and BRAZIUNAS, 1993) was applied prior to calibration. Calibrated age ranges were determined from the dendro-calibrated data of STUIVER and BECKER (1993) and STUIVER and BRAZIUNAS (1993) with the CALIB 3.0.3A program of STUIVER and REIMER (1993). The laboratory multiplier was set equal to 2 for all calibrations, and calibrated age ranges are expressed as a 68.3% confidence interval in yrs BP.

RESULTS

Stratigraphy

The total thickness of cored sediment ranges from 0.85 m near the lake outlet to 4.85 m at the most southerly site (Figure 3). The depth of penetration was limited by basal, intertidal sandy sediments.

The lowest unit in the sequence is composed of olive-gray (Munsell colour 5Y 4/2), fine to very fine sand, silty sand, and sandy mud containing shells of marine bivalves and, locally, wood. The finer facies of this unit are common only at the most southerly site; here the unit consists of interlayered sandy mud and muddy fine sand. At most other sites, the sand is clean and lacks conspicuous stratification.

The sand is overlain by yellowish-brown (10YR 3/2) gyttja. The contact between the sand and the gyttja is abrupt and rises nearly 3.5 m towards the lake outlet (Figure 3). In six of the seven cores the gyttja is structureless. At the most southerly core site, however, the lower part of the sequence contains two laminated zones 14 and 16 cm thick; individual laminae within these zones are 1–3 mm thick and are defined by differences in colour (brown to black). The gyttja unit thins to the north from about 4 m thick in the southern basin to 0.6 m at the lake outlet (Figure 3).

There are two conspicuous horizons in the gyttja sequence. A thin (3–10 mm) layer of olive-gray (5Y 3/2, 5Y 4/2) clay occurs near the middle of the gyttja sequence in the three most southerly cores (Figure 3). The basal contact of the clay layer is sharp; in contrast, the upper contact grades through a thin layer of organic-rich mud into the overlying gyttja. In each of these three cores the gyttja directly below the clay layer is unusually dark. The reason for this is not known. The second conspicuous horizon is a 9 cm thick mixture of plant detritus (branches, twigs, bryophyte fragments and conifer needles, seeds and cones) and sand in the lower part of the gyttja sequence in a core near the lake outlet (core 6; Figure 3). This layer has a sharp lower contact and a gradational upper contact. The same layer occurs in a core to the north (core 7; Figure 3), but is thicker (20 cm) and coarser there. At that site the plant debris coarsens and increases in abundance downward, inhibiting penetration by the Livingstone corer.

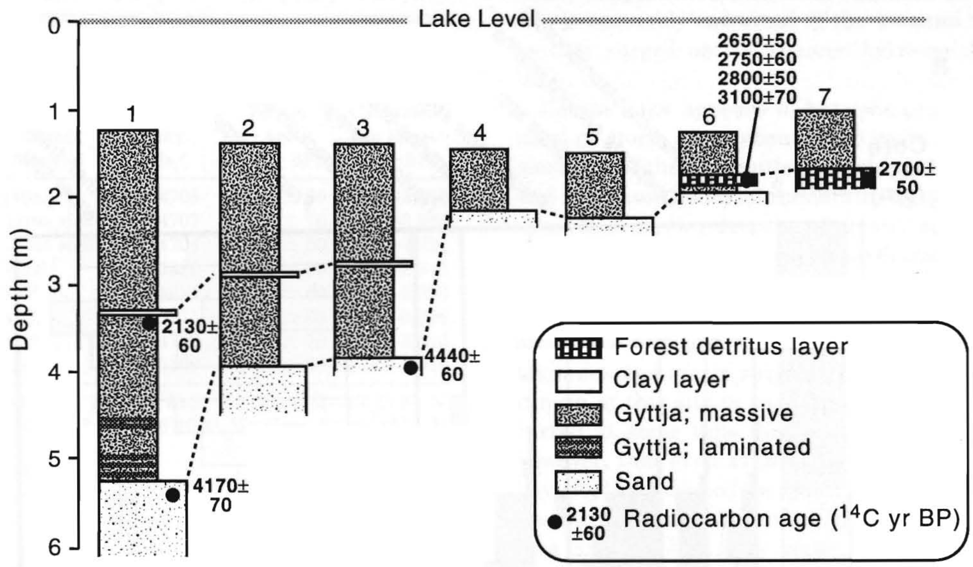

Figure 3. Stratigraphy of cores from Kanim Lake. Core locations shown in Figure 1d.

Diatom Biofacies

The diatom assemblage in the basal sand in core 2 is dominated by valves of marine and marine-brackish species (Figure 4a), providing further evidence for a marine origin of this unit. *Plagiogramma staurophorum* and *Opephora pacifica* and *O. marina,* which RAO and LEWIN (1976) noted as common members of the epipsammon in an intertidal sandflat in the Pacific Northwest, are particularly abundant in this basal sand assemblage, and indicate that the sea had open access to the basin at the time of deposition. The abundance of benthic species, and the fact that many of the valves are not broken or abraded, suggest that the basal sand unit is not a storm or tsunami deposit.

Gyttja overlying the basal sand contains a freshwater diatom community dominated by *Fragilaria* species. Just above the sand-gyttja contact, however, there are a few valves of marine diatoms, and valves of mesohalobous species such as *Navicula peregrina* and *Thalassiosira lacustris* are common, indicating a short-lived brackish phase transitional from a marine to a lacustrine environment.

Diatom assemblages associated with the clay layer were examined in core 1. Deposition of the clay layer coincided with small-scale and short-lived changes in the freshwater diatom community of the lake; *Cymbella minuta,* for example, exhibited a transient increase in abundance at the expense of other species such as *Aulacoseira granulata* (Figure 4b). No diatoms of a marine or brackish affinity, however, were noted in this layer or in the organic mud above it.

Diatom assemblages in the sandy forest detritus layer were determined from single samples recovered from cores 6 and 7 (Figure 4c). The diatom communities in both samples consist primarily of the same freshwater species found in the gyttja unit, but there are also substantial numbers of brackish diatoms (23% of the total in both samples) and some marine and marine-brackish forms (0.6% and 3% of the total in cores 6 and 7, respectively). As in the case of the gyttja just above the basal sand, the most abundant brackish diatom in the sandy forest detritus layer is *Navicula peregrina.* About two-thirds of the marine species in the forest detritus layer consist of planktonic forms, and most of these valves are broken or abraded, indicating transport prior to deposition. The diatom assemblage in the sandy detrital layer therefore appears to comprise material reworked from the lake bottom (the freshwater species), nearby intertidal marshes (brackish forms) and coastal and offshore sources (the marine forms).

Radiocarbon Ages

Whole shells and shell fragments collected from the upper part of the marine sand-mud unit in cores 1 and 3 yielded AMS radiocarbon ages of 4170 ± 70 and 4440 ± 60 ^{14}C yr BP (Table 1). Five AMS ages were obtained on cones and twigs from the layer of plant detritus and sand in cores 6 and 7 near the outlet. Four of the ages are between 2,650 and 2,800 ^{14}C yr BP; one sample yielded an age of 3,100 ± 70 ^{14}C yr BP. A thin slice of gyttja taken from directly below the clay layer in core 1 was dated at 2,130 ± 60 ^{14}C yr BP.

The calibrated ages of these samples (Table 1) indicate that the changeover from a marine to a freshwater lacustrine environment at Kanim Lake probably occurred 3,500 to 4,000 years ago. The layer of coarse plant detritus and sand in the vicinity of the lake outlet was deposited about 2,800 years ago. The thin clay layer was deposited between 1,900 and 2,300 years ago.

DISCUSSION

The calibrated radiocarbon ages cited above indicate that Kanim Lake emerged 3,500 to 4,000 years ago. Since emer-

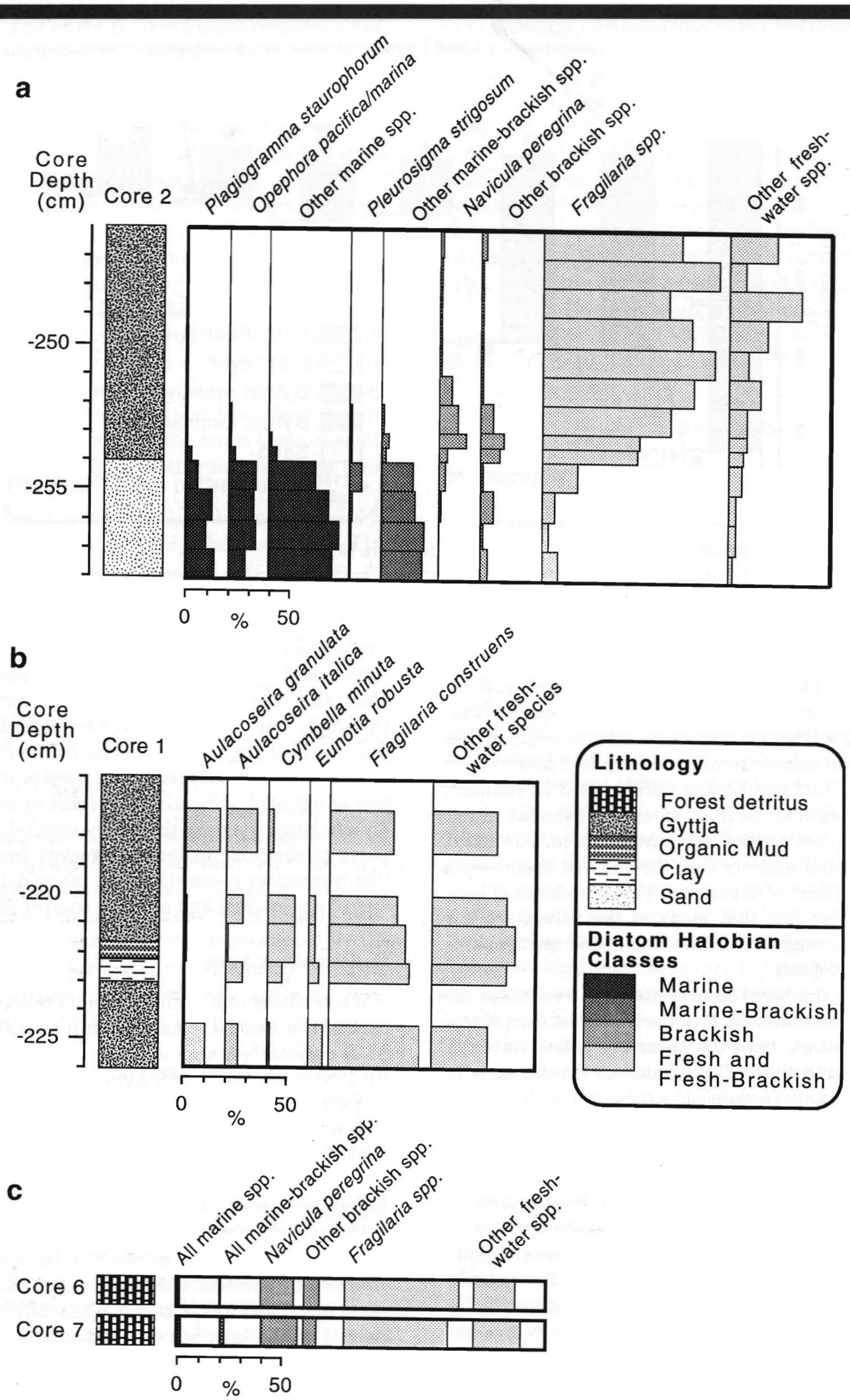

Figure 4. Representative diatom assemblages from (a) the basal sand and gyttja, (b) the clay layer, and (c) the sandy forest detritus layer.

Table 1. *Radiocarbon ages, Kanim Lake.*

Core[1]	Depth (m)[2]	Dated Material	Laboratory No.[3]	Conventional Radiocarbon Age (^{14}C yr BP)[4]	Calibrated Age Range (cal yr BP)[5]
1	2.13	Gyttja	TO-4708	2,130 ± 60	1,930–2,310
2	4.00–4.07	Marine shell	TO-4707	4,170 ± 70	3,520–3,880
3	2.59–2.94	Marine shell	TO-4704	4,440 ± 60	3,890–4,230
6	0.53–0.60	Twigs[6]	TO-5310	2,650 ± 50	2,720–2,840
7	0.53–0.60	Cone[7]	TO-4914	2,750 ± 60	2,750–3,000
6	0.53–0.60	Twig[8]	TO-5311	2,800 ± 50	2,780–3,020
6	0.53–0.60	Twig[8]	TO-4706	3,100 ± 70	3,060–3,440
7	0.63–0.83	Cone[9]	TO-5309	2,700 ± 50	2,740–2,920

[1]Locations: Core 1—49°23.5′ N, 126°20.2′ W; Core 3—49°23.6′ N, 126°20.3′ W; Core 6—49°23.7′ N, 126°20.2′ W; Core 7—49°23.7′ N, 126°20.2′ W

[2]Depth of sample below lake floor

[3]TO: IsoTrace (University of Toronto)

[4]Laboratory reported errors are $\pm 1\sigma$. All ages corrected to ^{13}C = −25‰ PDB

[5]Calibrated age ranges for plant material (TO-4708, TO-4914, and TO-4706) were determined from the dendro-calibrated decadal data of Stuiver and Becker (1993). Calibrated age ranges for marine shell fragments (TO-4707 and TO-4704) were determined from the dendro-calibrated bidecadal data of Stuiver and Braziunas (1993), with ΔR set to 390 ± 25 years. The ranges represent the 68.3% confidence interval based on the 2σ error limits of the radiocarbon age (error multiplier = 2)

[6]*Myrica gale*? (sweet gale?)

[7]*Thuja plicata* (western red cedar)

[8]*Picea sitchensis* (Sitka spruce)

[9]*Tsuga heterophylla* (western hemlock)

gence, about 4.5 m of gyttja has accumulated in the deepest part of the southern lake basin, and about 1 m of gyttja has accumulated near the lake outlet. Gyttja accumulation in the lake basin appears to have been interrupted on only two occasions: about 2,800 years ago a layer of sand and forest detritus was deposited near the lake outlet, and 1,900 to 2,300 years ago a thin layer of clay was deposited in the southern lake basin.

Several lines of evidence suggest that the sandy forest detritus layer was deposited by a tsunami or storm surge and not by a flood or some other process. No creeks enter the lake in the vicinity of cores 6 and 7; this and the presence of marine diatoms shows that the forest detritus layer is not a flood or slopewash deposit. The layer is thickest and most distinct near the lake outlet, which is the area where a tsunami or storm surge would enter the lake, and where most of the suspended and tractive load would be deposited. The sand component of this layer is likely derived from the beach, from shallow offshore sources, or from the outlet channel. The presence of planktonic marine diatoms is further evidence that at least some of this material came from offshore. Brackish-water diatoms account for almost one-quarter of the diatom assemblage in the forest detritus layer, implying that brackish marshes existed near the lake at that time, perhaps behind a barrier beach at the seaward end of the outlet stream. The plant detritus consists of wood and bryophyte fragments, and conifer needles, cones and seeds in various stages of decomposition. This is typical forest floor material in the temperate rain forest of western Vancouver Island and was presumably entrained in the tsunami or in storm waves as they surged up the channel between the ocean and the lake.

Kanim Lake appears to bear the imprint of a single tsunami or storm event about 2,800 years ago. At that time the elevation of the lake outlet was close to 3.5 m asl (Figure 2), and the waves had a minimum run-up of 1.5 m. The apparent absence of similar deposits of greater age may simply reflect our inability to distinguish these deposits from sandy subtidal and intertidal sediments that accumulated in the basin before it emerged. The absence of younger deposits may be due to the fact that the lake rose above the run-up limit for tsunamis and storm surges. This would limit the maximum run-up at this site to ca. 2 m, assuming that a high tide occurred at some time during the event. It should be noted, however, that even as the lake basin rose, other factors may have further reduced the chances of a tsunami or storm surge reaching the lake. In particular, the forest zone between the lake and the sea grew wider as the lake emerged; dense stands of trees reduce wave energy, and may inhibit inflow to a lake still within the elevational range of the run-up zone of a tsunami or surge on the open coast. The development of a belt of forest nearly 1 km wide along the outlet stream between Kanim Lake and the Pacific Ocean since emergence has undoubtedly been a major hindrance to tsunami access to this lake basin.

Although in the discussion so far both tsunamis and storms have been considered equally likely causes of the sandy forest detritus layer, we favor a tsunami origin. The emplacement of this deposit by storm waves or a storm surge is considered unlikely, for the following reasons. First, the mouth of the outlet stream from Kanim Lake lies on the relatively protected eastern shore of Hesquiat Harbour (Figure 1c) and the maximum shore-normal fetch in the vicinity of the stream mouth is 10 km. Waves generated across this limited fetch are unlikely to cause flow reversal in the outlet channel and deposition of marine material in Kanim Lake. Second, storm surges generated by onshore winds from the open Pacific are of limited magnitude because the waters overlying the narrow continental shelf off western Vancouver Island are relatively deep. The Tofino (Figure 1b) tide gage record, for example, indicates a maximum super-elevation of the water surface of 0.9 m during storms in the period 1929–1991 (Figure 5). Extrapolating from this record, we infer that storm surges equivalent to the >1.5 m event that breached Kanim Lake 2,800 years ago may have a recurrence interval exceeding 1,000 years (Fig. 5). This is substantially greater than the estimated average recurrence interval of interplate earthquakes on the Cascadia subduction zone, which is about 500 years (Atwater and Hemphill-Haley, 1996).

The third reason for discounting the storm hypothesis is that the sandy detritus layer may be the same age as one of the Cascadia plate-boundary earthquakes inferred from buried peat sequences in estuaries in Oregon (Nelson, 1992; Darienzo and Peterson, 1995; Nelson *et al.* 1996) and southern Washington (Atwater, 1988; Atwater and Hemphill-Haley, 1996). Cappings of tsunami sand lie above this buried peat in several of these estuaries. Radiocarbon ages

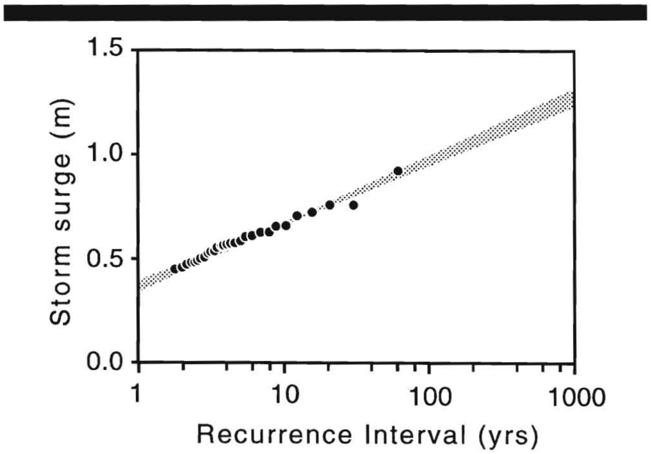

Figure 5. Recurrence interval-magnitude relations for storm surges on the west coast of Vancouver Island. Storm surge magnitude was determined from readings of water surface super-elevation (= observed − predicted tide height) at the Tofino tide gage between 1929–1973 and 1976–1991. Recurrence intervals (RI) for individual events were calculated by means of the Weibull equation: $RI = (n + 1)/r$, where n is the length of the record and r is the rank of the event in the series. The 95% confidence interval for the slope of the RI-storm magnitude linear regression equation is shown by the shaded band. Data supplied by F. Stephenson, Institute of Ocean Sciences, Sidney, British Columbia.

for this earthquake, reported as "event 5" by ATWATER (1988) and "event 6" by DARIENZO and PETERSON (1995), range from 2,500 to 3,220 ^{14}C yr BP; the overall average is 2,800 ^{14}C yr BP.

We do not consider the thin clay layer in the southern basin to be a tsunami deposit. The clay layer was not found in cores near the lake outlet where one might expect to find it if it were deposited by a tsunami. The exclusively freshwater character of the associated diatom community (Figure 4b) also argues for a non-tsunami origin. The clay layer is most likely the product of inwash of sediment into the southern lake basin from the slopes to the east, possibly due to a severe storm or to a landslide.

CONCLUSIONS

Most geological research on tsunamis has been done in intertidal marshes where the challenge to the investigator is to discriminate between tsunami deposits and those resulting from storm surges and floods. On emerging coastlines this research is further constrained by the short lifespan of marshes. In these settings nearshore lakes offer several advantages for paleotsunami investigation. These include the distinctive lithological and ecological character of tsunami deposits, the high preservation potential of such deposits, and the relative ease of dating them. Kanim Lake illustrates these advantages, as well as some of the limitations of using lake basins as sources of information about tsunami magnitude and recurrence. Since its emergence from the sea 3,500–4,000 years ago, the basin of Kanim Lake has risen at a rate of about 1 m ka^{-1}. Although the lake still lies below the potential run-up limit for tsunamis generated by earthquakes on the northern Cascadia subduction zone, only one tsunami has apparently entered the lake since emergence. This event occurred about 2,800 years ago and had a minimum amplitude of 1.5 m. The absence of a record of more recent events can be partly explained by the decreasing probability of tsunami penetration as the lake rose through the potential run--up zone. A more important limitation on tsunami access to the site over this period, however, has been the progressive widening of the intervening land between the ocean and the lake and the development of dense forest stands on this surface.

In closing, we suggest that paleotsunami researchers working on emerging coastlines not restrict their investigations to single sites, but rather attempt to sample sites at a variety of elevations in order to better assess tsunami run-up and recurrence. They should also recognize the problems associated with increases in the distance of sites from the sea through time on emergent coastlines, particularly in environments where dense forests develop rapidly.

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LITERATURE CITED

ATWATER, B.F., 1987. Evidence for great Holocene earthquakes along the outer coast of Washington. *Science,* 236, 101–103.

ATWATER, B.F., 1988. Geologic studies for seismic zonation of the Puget lowland. *In:* Jacobsen, M.L. and Rodriguez, T.R., compilers, National Earthquake Hazards Reduction Program, summaries of technical reports, Volume 25. *U.S. Geological Survey Open-File Report 88–16,* pp. 120–133.

ATWATER, B.F., 1992. Geologic evidence for earthquakes during the past 2000 years along the Copalis River, southern coastal Washington. *Journal of Geophysical Research,* 97, 1901–1919.

ATWATER, B.F. and HEMPHILL-HALEY, E. 1996. *Preliminary Estimates of Recurrence Inetrvals for Great Earthquakes of the Past 3500 Years at Northeastern Willapa Bay, Washington.* U.S. Geological Survey Open-File Report 96–001, 88 p.

ATWATER, B.F. and MOORE, A.L., 1992. A tsunami about 1000 years ago in Puget Sound, Washington. *Science,* 258, 614–617.

ATWATER, B.F.; NELSON, A.R.; CLAGUE, J.J.; CARVER, G.A.; YAMAGUCHI, D.K.; BOBROWSKY, P.T.; BOURGEOIS, J.; DARIENZO, M.E.; GRANT, W.C.; HEMPHILL-HALEY, E.; KELSEY, H.M.; JACOBY, G.C.; NISHENKO, S.P.; PALMER, S.P.; PETERSON, C.D., and REINHART, M.A., 1995. Summary of coastal geological evidence for past great earthquakes at the Cascadia subduction zone. *Earthquake Spectra,* 11, 1–18.

CLAGUE, J.J., 1981. *Late Quaternary Geology and Geochronology of British Columbia. Part 2: Summary and Discussion of Radiocarbon-Dated History.* Geological Survey of Canada Paper, 80–35, 41p.

CLAGUE, J.J. and BOBROWSKY, P.T., 1994a. Evidence for a large earthquake and tsunami 100–400 years ago on western Vancouver Island, British Columbia. *Quaternary Research,* 41, 176–184.

CLAGUE, J.J. and BOBROWSKY, P.T., 1994b. Tsunami deposits beneath tidal marshes on Vancouver Island, British Columbia. *Geological Society of America Bulletin,* 106, 1293–1303.

CLAGUE, J.; HARPER, J.R.; HEBDA, R.J., and HOWES, D.E., 1982.

Late Quaternary sea levels and crustal movements, coastal British Columbia. *Canadian Journal of Earth Sciences,* 19, 597–618.

DARIENZO, M.E. and PETERSON, C.D., 1990. Episodic tectonic subsidence of late Holocene salt marshes, northern Oregon central Cascadia margin. *Tectonics,* 9, 1–22.

DARIENZO, M.E. and PETERSON, C.D., 1995. Magnitude and frequency of subduction-zone earthquakes along the northern Oregon coast in the past 3,000 years. *Oregon Geology,* 57, 3–12.

DARIENZO, M.E.; PETERSON, C.D., and CLOUGH, C. 1994. Stratigraphic evidence for great subduction zone earthquakes at four estuaries in northern Oregon. *Journal of Coastal Research,* 10, 850–876.

DAWSON, A.G.; FOSTER, I.D.L.; SHI, S.; SMITH, D.E., and LONG, D., 1991. The identification of tsunami deposits in coastal sediment sequences. *Science of Tsunami Hazards,* 9, 73–82.

FOGED, N., 1981. *Diatoms in Alaska.* Vaduz: J. Cramer, Bibliotheca Phycologia, 317 p.

FRIELE, P.A. and HUTCHINSON, I., 1993. Holocene sea-level change on the central-west coast of Vancouver Island, British Columbia. *Canadian Journal of Earth Sciences,* 30, 832–840.

HAWORTH, E.Y., 1976. Two Late-Glacial (Late Devensian) diatom assemblage profiles from northern Scotland. *New Phytologist,* 77, 227–256.

HEMPHILL-HALEY, E., 1993. *Taxonomy of Recent and Fossil (Holocene) Diatoms (Bacillariophyta) from Northern Willapa Bay, Washington.* United States Geological Survey Open-File Report, 93-289. 151 p.

HEMPHILL-HALEY, E., 1995. Diatom evidence for earthquake-induced subsidence and tsunami 300 yr ago in southern coastal Washington. *Geological Society of America Bulletin,* 107, 367–378.

HENDEY, N.I., 1964. *An Introductory Account of the Smaller Algae of British Coastal Waters. Part V: Bacillariophyceae (Diatoms).* London: Ministry of Agriculture, Fisheries and Food, Fisheries Investigations Series IV, 317p.

HUSTEDT, F., 1953. Die Systematik der Diatomeen in ihren Beziehungen zur Geologie und ökologie nebst einer Revision des Halobien-Systems. *Svensk botanisk Tidskrift,* 47, 509–519.

HUTCHINSON, I., 1992. *Holocene Sea-Level Change in the Pacific Northwest: a Catalogue of Radiocarbon Dates and an Atlas of Regional Sea-Level Curves.* Simon Fraser University, Institute of Quaternary Research, Discussion Paper No. 1, 100p.

HUTCHINSON, I.; ROBERTS, M.C., and WILLIAMS, H.F.L., 1995. Stratigraphy, diatom biofacies, and palaeogeomorphology of a mid Holocene distributary channel system, Fraser River delta, British Columbia, Canada. *Canadian Journal of Earth Sciences,* 32, 749–757.

HYNDMAN, R.D., 1995. Giant earthquakes along the west coast of North America. *Scientific American,* 273, 68–75.

KANAMORI, H. and KICKUCHI M., 1993. The 1992 Nicaragua earthquake: a slow tsunami earthquake associated with subducted sediments. *Nature,* 361, 714–716.

KOLBE, R.W., 1927. Zur ökologie, Morphologie und Systematik der Brackwasser-Diatomeen. *Pflanzenforschung,* (Jena: Gustav Fischer), 7, 146p.

LAWS, R.A., 1988. Diatoms (Bacillariophyceae) from surface sediments in the San Francisco Bay estuary. *Proceedings of the California Academy of Sciences,* 45, 133–254.

LONG, D.; SMITH, D.E., and DAWSON, A.G., 1989. A Holocene tsunami deposit in eastern Scotland. *Journal of Quaternary Science,* 4, 61–66.

MINOURA, K. and NAKAYA, S., 1991. Traces of tsunami preserved in inter-tidal lacustrine and marsh deposits: some examples from northeast Japan. *Journal of Geology,* 99, 265–287.

NELSON, A.R., 1992. Discordant ^{14}C ages from buried tidal-marsh soils in the Cascadia subduction zone, southern Oregon coast. *Quaternary Research,* 38, 74–90.

NELSON, A.R. and KASHIMA, K., 1993. Diatom zonation in southern Oregon tidal marshes relative to vascular plants, foraminifera, and sea level. *Journal of Coastal Research,* 9, 673–697.

NELSON, A.R., JENNINGS, A.E., and KASHIMA, K., 1996. An earthquake history derived from stratigraphic and microfossil evidence of relative sea-level change at Coos Bay, southern coastal Oregon. *Geological Society of America Bulletin,* 108, in press.

NG, M.K.-F.; LEBLOND, P.H., and MURTY, T.S., 1990. Simulation of tsunamis from great earthquakes on the Cascadia subduction zone. *Science,* 250, 1248–1251.

PATRICK, R. and REIMER, C.W., 1966. *The Diatoms of the United States—Exclusive of Alaska and Hawaii.* Philadelphia: Monographs of the Academy of Natural Sciences, 13 (1), 688 p.

PATRICK, R. and REIMER, C.W., 1975. *The Diatoms of the United States—Exclusive of Alaska and Hawaii.* Philadelphia: Monographs of the Academy of Natural Sciences, 13 (2, pt.A), 213p.

PIENITZ, R.; LORTIE, G.; and ALLARD, M., 1991. Isolation of lacustrine basins and marine regression in the Kuujjuaq area, northern Québec, as inferred from diatom analysis. *Géographie physique et Quaternaire,* 45, 155–174.

RAO, V.N.R. and LEWIN, J., 1976. Benthic marine diatom flora of False Bay, San Juan Island, Washington. *Syesis,* 9, 173–213.

REINHART, M.A. and BOURGEOIS, J., 1987. Distribution of anomalous sand at Willapa Bay, Washington—evidence for large-scale landward-directed processes. *Eos,* 68, 1469.

STUIVER, M. and BECKER, B., 1993. High-precision decadal calibration of the radiocarbon time scale, AD 1950–6000 BC. *Radiocarbon,* 35, 35–65.

STUIVER, M. and BRAZIUNAS, T.F., 1993. Modeling atmospheric ^{14}C influences and ^{14}C ages of marine samples to 10,000 BC. *Radiocarbon,* 35, 137–190.

STUIVER, M. and REIMER, P.J., 1993. Extended ^{14}C data base and revised CALIB 3.0 ^{14}C age calibration program. *Radiocarbon,* 35, 215–230.

VAN DER WERFF, A. and HULS, H., 1957–1974. *Diatomeeënflora van Nederland.* Den Haag: Abcoude.

VOS, P.C. and DE WOLF, H., 1988. Methodological aspects of palaeoecological diatom research in coastal areas of the Netherlands. *Geologie en Mijnbouw,* 67, 31–40.

WHITMORE, P.M., 1993. Expected tsunami amplitudes and currrents along the North American coast for Cascadia subduction zone earthquakes. *Natural Hazards,* 8, 59–73.