Geomorphology and accommodation space as limiting factors on tsunami deposition: Chatham Island, southwest Pacific Ocean

S.L. Nichol a,⁎, C. Chagué-Goff b, J.R. Goff b, M. Horrocks a,c, B.G. McFadgen d, L.C. Strotz b

a School of Geography and Environmental Science, University of Auckland, Private Bag 92019, Auckland, New Zealand
b Natural Hazards Research Lab and Australian Tsunami Research Centre, School of Biological, Environmental and Earth Sciences, University of New South Wales, Sydney, NSW 2052, Australia
c Microfossil Research Ltd., 31 Mont Le Grand Rd., Mt Eden, Auckland, New Zealand
d School of Maori Studies, Victoria University of Wellington, PO Box 600, Wellington, New Zealand

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A B S T R A C T

Chatham Island in the southwest Pacific Ocean is exposed on all sides to potential tsunami impact. In historical time, tsunamis are known to have inundated the coast on several occasions, with the largest event in 1868. Coastal dunes along the northeast coast of Chatham Island preserve sedimentary evidence of this and possibly earlier tsunami events, as localised gravel lags. However, these deposits lack a clear stratigraphic context and establishing their age is difficult. This study examines the sediment record in a freshwater wetland at Okawa Point, located directly landward of the dunes where apparent tsunami gravels occur. Sediment descriptions, pollen, foraminifera, chemical data and radiocarbon dates from cores are used to reconstruct the environmental history of the wetland. The record extends from ca. 43 ka to the present and incorporates glacial, post-glacial and human-influenced phases. Throughout this time the wetland appears to have remained isolated from catastrophic marine inundation. The only evidence for saltwater intrusion is observed in the historic period, via geochemical, grain size and pollen data, which record a marine inundation event that forced the transport of a thin (cm-thick) deposit of dune and beach sand into the seaward edge of the wetland. This is interpreted as the signature of the 1868 tsunami. The lack of more widespread physical evidence for this and other tsunami events in the wetland is attributed to the morphological roughness afforded by coastal dunes and limited accommodation space for Holocene deposits.

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1. Introduction

The preservation of tsunami deposits in the coastal sedimentary record is primarily a function of the morphology of the pre-tsunami landscape and the available accommodation space. Estuaries, lagoons and wetlands located behind low barriers have proven to be excellent repositories of tsunami deposits along source-bordering coasts, notably the US Pacific Northwest (Peters et al., 2007), Chile (Atwater et al., 1992; Cisternas et al., 2005), Japan (Minoura and Nakaya, 1990, Minoura et al., 1994) and New Zealand (Goff et al., 2001). In situations where tsunami inundation is associated with co-seismic subsidence of the receiving basin, the preservation potential of deposits is further enhanced to the extent that a history of multiple events spanning thousands of years can be reconstructed (e.g. Atwater, 1987).

The potential to extract a history of tsunami inundation from coastal sediments is not limited to source-bordering coasts. For example, the Pacific-wide tsunami generated by the AD 1960 Chile earthquake deposited a sand sheet across beach ridges at the head of Miyako Bay, northeast Japan (Onuki et al., 1961). A similar trans-Pacific link has been proposed by Atwater et al. (2005) between the AD 1700 Cascadia earthquake (northwest USA) and widespread tsunami damage in eastern Japan, although tsunami deposits from this event have yet to be documented (Pararas-Carayannis, 2006). Considerable scope remains therefore to explore the sedimentary record for tsunami evidence at localities that are not source-bordering but are situated within the path of known tsunamis.

At distal locations such as mid-ocean islands the degree of tsunami impact can be highly spatially variable and is largely determined by the width and gradient of the submarine shelf around an island, as a control on tsunami wave height, and by the coastal geomorphology of an island (Kench et al., 2008). Coastal dunes in particular can afford significant protection to tsunami inundation, as observed on the coast of Sri Lanka following the 2004 Indian Ocean tsunami (Liu et al., 2005; Morton et al., 2008). Chatham Island, in the southwest Pacific Ocean, is at a distal location relative to the main tsunami sources in the ocean basin. The island has been inundated by several tsunamis during historical times (de Lange and Healy, 1986), the largest on 15 August 1868, when a submarine earthquake off the coast of Peru generated a series of at least three large waves that crossed the northern, eastern and western coasts of the island. Eyewitness accounts of the event...
reported one death and significant property damage, deposition of sand and marine debris across high ground and flooding of lowland areas (Anonymous, 1868; Hector, 1868). To date, however, the physical record of that event remains undiscovered. Here we use the sediment record from a dune and wetland system at Okawa Point on the exposed northeast coast of Chatham Island to: (i) reconstruct the environmental history of the wetland; (ii) test for sedimentary evidence of the 1868 tsunami, and; (iii) evaluate that evidence in the context of the geomorphic setting of the wetland.

2. Chatham Island

Chatham Island is located 850 km to the east of the South Island of New Zealand at 44° S, 176° 30' W at the eastern end of Chatham Rise (Fig. 1). Covering an area of 910 km², the island is surrounded by a shelf that is 12–50 km wide with a shelf break at 150–200 m water depth. This location and bathymetric setting exposes Chatham Island to potential impact from tsunamis that may be focused and amplified across Chatham Rise, with potential source areas at far-field localities (e.g. South America subduction zone) and regionally, including the subduction zone at the boundary between the Australian and Pacific Plates (beneath New Zealand) and submarine volcanoes on the Kermadec Ridge (Goff et al., 2010b).

The island has mixed geological origins, incorporating: the eroded remnants of a Late Cretaceous strato-volcano that forms the southern and highest part of the island, rising to an elevation of 294 m; sedimentary rocks of Late Cretaceous and Cenozoic age that underlie the undulating terrain of the central area at 20–60 m elevation; up-thrown outcrops of metamorphic basement rocks (Chatham Schist) along the northern coast that form headlands and shore platforms, and; inactive volcanic cones of late Eocene to Oligocene age on the northern part of the island that range in elevation from 100 to 190 m (Hay et al. 1970; Grindley et al., 1977; Campbell et al., 1993; Adams et al., 2008). Despite this history of emergence and volcanism, the island has remained tectonically stable since the Late Cretaceous (Campbell et al., 1993).

Fig. 1. A. Locality map for Chatham Island in the western Pacific Ocean showing major bathymetric features, including Chatham Rise. B. Bathymetric map of the shelf surrounding Chatham Island and Pitt Island (adapted from Mackay et al., 2004).
Landforms of Quaternary age on Chatham Island include lakes, wetlands, dunes and coastal sand barriers that have formed principally along the north and east coasts, and enclose Te Whanga Lagoon, the largest geomorphic feature on the island, that covers 180 km² (20% of the island). Holocene coastal dunes comprise two to three sub-parallel ridges that follow the alignment of the modern coast. Surface soils, buried soils and pumice deposits exposed in coastal dune sections allowed McFadgen (1994) to propose four depositional episodes, as follows: the Te Onean Depositional Episode between ca. 5000 and 2200 yrs BP; the Okawan Depositional Episode between 2200 and 450 yrs BP; the Kekerionean Depositional Episode between ca. 450 and 150 yrs BP, and; the Waitangian Depositional Episode between 150 yrs BP and the present. These episodes were punctuated by periods of relative stability of dunes, during which time soils formed under vegetation cover (McFadgen, 1994). In places these dune sequences are stacked vertically reaching to 9 m in height.

Based on correlation of archaeological sites using dated dune soils as chronological markers, McFadgen (1994) also proposed that the timing of human settlement of Chatham Island by the Moriori (Maori Polynesians) was between 450 and 400 yrs BP. Periods of dune instability were therefore unrelated to human activity. Rather, it was suggested that each depositional episode followed storms that caused reworking of coastal sands via longshore currents and in turn by wind to supply dunes (McFadgen, 1994).

3. Study site

Okawa wetland (43° 45.5′ S, 176° 16.42′ W) is located on the northeast coast of Chatham Island, 2.5 km to the west of Okawa Point (Fig. 2). The wetland extends 1.1 km along the coast behind Holocene foredunes and is approximately 100 m wide with fixed Pleistocene dunes on the landward side. The Holocene sand dunes are up to 8 m high and 120–150 m wide, forming a continuous barrier. The only connection to the open coast is through a narrow channel, less than 3 m wide that has cut through the dunes to the beach. During fieldwork in January 2007 this channel was blocked by a sandy beach berm at its seaward end. Also at this time, bedrock (Chatham Schist) was exposed on the lower beach face forming a discontinuous shore platform (Fig. 3a). The wetland surface is mostly vegetated by grasses and sedges, with the exception of two small ephemeral lakes located at the western end and the central section of the wetland. The hydrology of the wetland is supplied chiefly by direct rainfall and

![Fig. 2. A. Map of Chatham Island showing areas of coastline that were inundated by the 1868 tsunami, indicated by black shading (based on de Lange and McSaveney, 2009). B. Photograph of Okawa wetland and dunes looking southeast, C. Geomorphic map of Okawa wetland showing locations of core sites and pebble lag deposit in foredunes.](image-url)
groundwater flow from a catchment area of ~0.8 km², as there are no major streams or channels feeding the wetland.

McFadgen (1994) described a deposit of sea-rafted Loisels pumice exposed in a section of eroded foredune at the eastern end of the beach toward Okawa Point. Based on observed deposits of flotsam on the backshore, this deposit is at an equivalent elevation attained by modern storm surge. This suggests that storm surges do not enter the channel through the dunes, nor the wetland. This observation is based on the modern geomorphology of the dunes and does not preclude tidal or storm inundation of the wetland during earlier stages of its development. At Okawa Point west, McFadgen (1994) mapped the Waitangian Depositional Episode (~150 yrs BP), comprising grey sands of the surface dunes overlying a buried palaeosol of unknown age that relates to either the Okawan (2200–450 yrs BP) or Kekerionean (450–150 yrs BP) Depositional Episodes. The Okawan/Kekerionean dune episode is also mapped by McFadgen (1994) as a surface deposit that forms a sand sheet along the seaward edge of Okawa wetland (Fig. 2). On this basis, it is inferred that the Okawa wetland has been enclosed by dunes for the past 2200 years, possibly longer.

The dunes at Okawa also feature surface lag deposits comprising gravel to pebble-sized clasts of schist mixed with shell and (whale?) bone fragments (Fig. 3b). These deposits occur as discontinuous scatters within deflation hollows in the dunes, and up to a maximum elevation of 10 m above mean sea level. At one site, the deposit includes a boulder sized slab of Chatham Schist that most likely comes from the shore platform. In all cases the pebbles and gravels are well to very well rounded, which also suggests a shallow marine source. As a surface lag, these deposits lack a clear stratigraphic context; however, they were most likely deposited during the Waitangian Depositional Episode and are therefore less than 150 years in age.

An annotated map of Chatham Island depicting where the 1868 tsunami inundated the coast provides information to help interpret
the age and origin of the dune gravels at Okawa (de Lange and McSaveney, 2009). This map shows that all of the low-lying land at Okawa Point and some of the adjacent shoreline were inundated (Fig. 2). A note from the map's creator, Thomas Ritchie, states that the average rise in water level (run-up height) was approximately 20 ft (~7 m) above spring high tide level. For Okawa wetland, this almost certainly would have resulted in flooding to a depth of between one and three metres. It is also highly likely that the dunes at Okawa were inundated by the 1868 tsunami. If so, the tsunami would have transported sediment from the nearshore and beach face, including the gravel and pebble material that now forms the lag deposit. Whether the tsunami also deposited sediment in the wetland is part of our investigation.

4. Methods

Four sediment cores (cores Ok1 to Ok4) were collected from Okawa Point wetland using a vibracoring system. Core penetration ranged from 1.5 m to 3.6 m, with three cores meeting refusal on compact dry sand and gravel (cores Ok1, Ok2) and rock (core Ok3). Compaction of each core sample was measured prior to retrieval and ranged from 6 cm to 32 cm. The elevation of core sites was measured using a Garmin differential GPS relative to high tide mark on the beach and is estimated to be accurate to within 0.5 m (Table 1).

In the laboratory, cores were split lengthwise and logged to describe sediment texture and sedimentsary structures. Core Ok3 was selected as representative of the wetland stratigraphy and was subsampled for pollen, foraminifera, and diatom analyses, measurement of organic content and for radiocarbon dating. Sampling depths are shown on Fig. 5. Due to the high organic content of core Ok3 it was considered inappropriate to conduct grain size measurements (i.e. laser granulometry). Therefore, no samples were taken for this purpose and all sediment descriptions are based on visual classification (Folk, 1974). For pollen analysis, 25 sediment samples were prepared following the standard acetylation and hydrofluoric acid method (Moore et al., 1991), with the hydrofluoric acid step replaced by density separation with sodium polytungstate (specific gravity 2.0). At least 250 pollen grains were counted for each sample, except the samples from 260 cm depth (177 counted) and 70 cm depth (211 counted). Three samples from the uppermost facies in core Ok3 (3, 7, and 10 cm depth) were analysed for foraminiferal content. Foram tests were separated from the sand residue by floating in sodium polytungstate solution, then picked, identified and counted using a binocular microscope. Diatom content was evaluated separately by microscope scanning of sediment smear slides for eight samples taken from different sediment facies in core Ok3. None of these slides contained diatom material, so no further analysis was undertaken.

Organic content was measured on 23 samples by loss-on-ignition (430 °C for 24 h) following the procedure of Gale and Hoare (1991). Three relatively sand-rich samples were selected for photomicroscopy of the sieved sand fraction to characterise grain shape and mineralogy in discrete sand beds. For radiocarbon dating, seven samples of organic-rich sediment (sandy peat and organic silt) were taken from core Ok3 and submitted to the University of Waikato Radiocarbon Dating Laboratory for AMS assay. These samples represented the main sediment facies and were positioned at key stratigraphic surfaces (Table 2). Conventional radiocarbon ages younger than 20 ka were calibrated using WInsCal08 against pine-cedar calibration data (Hogg et al., 2002) and OxCal 4.1 (Bronk Ramsey, 2009) for older ages (Table 2).

A second stage of laboratory work was carried out on core Ok1. The objective was to test for evidence of seawater incursion into Okawa wetland. Core Ok1 was chosen because of its proximity to the narrow stream outlet through the dunes. Geochemical analysis focused on the upper 50 cm of core Ok1 because this interval included a sand bed of potential marine origin within the peat. Nine subsamples were taken and split, with one part used to determine organic content by loss-on-ignition (as above) and the remaining material used for geochemical analysis. The latter split was oven-dried at 60 °C then ground, dried at 110 °C and analysed for elemental composition (sulphur, chlorine, bromine, sodium, calcium, strontium, magnesium and iron) by X-Ray Fluorescence at SpectraChem Analytical, Lower Hutt, New Zealand.

5. Results

5.1. Sediment facies

Five sediment facies are defined for the Okawa wetland on the basis of sediment texture, organic content and stratigraphic position (Figs. 4, 5). In core Ok3, four of these facies (A–D) are preserved and are described in detail below. The other cores recovered successions of two (core Ok4) or three (cores Ok2, Ok1) of these four facies. The fifth facies (Q) was recovered only in core Ok1.

Facies A is the basal deposit in the wetland that comprises a very poorly sorted mixture of angular gravel, coarse sand and silt (Fig. 5). In core Ok3, this facies is 66 cm thick, massive and rests directly on a hard rock surface at 3.1 m depth. Gravel clasts comprise weathered schist up to 4 cm in length and are mostly platy to rod-shaped. A photomicrograph of the sand fraction from facies A shows moderately sorted, very coarse grains of quartz and lithics that are sub-angular to sub-rounded (Fig. 5). Gravel is absent within the upper 15 cm of this facies, where silty coarse sand forms a massive interbed. Organic content within the gravely part of the deposit ranges between 12% and 16% and incorporates very coarse grains of quartz and lithics that are sub-angular to sub-rounded (Fig. 5). Geochemical analysis focused on the gravelly part of core Ok3, this facies is 66 cm thick, massive and rests directly on a hard rock surface at 3.1 m depth. Gravel clasts comprise weathered schist up to 4 cm in length and are mostly platy to rod-shaped. A photomicrograph of the sand fraction from facies A shows moderately sorted, very coarse grains of quartz and lithics that are sub-angular to sub-rounded (Fig. 5).

Table 1

<table>
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<th>Core ID</th>
<th>Easting</th>
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<th>Elevation above HW (m)</th>
<th>Penetration (cm)</th>
<th>Compaction (cm)</th>
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<td>676599</td>
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<td>152</td>
<td>10</td>
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<tr>
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<td>676509</td>
<td>6</td>
<td>359</td>
<td>32</td>
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<tr>
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<td>676211</td>
<td>5</td>
<td>341</td>
<td>29</td>
</tr>
<tr>
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<td>368740</td>
<td>676565</td>
<td>3</td>
<td>161</td>
<td>6</td>
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Table 2

<table>
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<tr>
<td>-------------------</td>
</tr>
<tr>
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<tr>
<td>74–73</td>
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<td>119–120</td>
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<td>145–146</td>
</tr>
<tr>
<td>195–196</td>
</tr>
<tr>
<td>245–246</td>
</tr>
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</table>

¹Calibrated age expressed at the 94.9% confidence interval, calculated using Winscal08 against the Waikato University pine-cedar calibration data (Hogg et al., 2002) smoothed using the method of Knox and McFadgen (2001).
²Calibrated ages are expressed at the 95.1% confidence interval, calculated using WInsCal08 against southern hemisphere atmospheric data from McCormac et al. (2004).
³Calibrated ages are expressed at the 94.9% confidence interval, calculated using WInsCal08 against southern hemisphere atmospheric data from McCormac et al. (2004).
⁴Calibrated ages are expressed at the 95.1% confidence interval, calculated using OxCal 4.1 against IntCal09 (Bronk Ramsey, 2009).
⁵Beyond the range of the calibration curve.
lenticular bed at 153 cm depth. Three samples of organic silt from 245 cm, 195 cm and 145 cm depth in facies B yielded conventional radiocarbon ages of >42,980 (Wk-21605), 41,695±2019 (Wk-21604) and 37,209±1108 (Wk-21603) yrs BP, respectively (Table 2; Fig. 5).

The contact between Facies B and C in core Ok3 is a sharp surface defined by an abrupt transition from peat to sand at 143 cm depth (Fig. 5). Facies C comprises a set of four silty sand beds with a combined thickness of 24 cm. The beds are distinguished by minor differences in grain size within the medium to coarse sand range and sharp basal contacts between each bed. In addition, a one centimetre-thick silt flaser bed separates the two upper sand beds. Organic material is present throughout the facies as silt-sized fragments forming mm-thick laminae, but comprising less than one percent of the sediment mass. A photomicrograph of the lowermost bed shows the sand fraction comprises well sorted, sub-rounded to rounded medium-sized (0.25–0.5 mm) quartz grains (Fig. 5). Facies C was not recovered in any of the other three cores, suggesting that it has limited spatial extent within the wetland.

Facies D is 120 cm thick in core Ok3. It forms the uppermost deposit within all cores but its thickness varies (Fig. 4). This facies comprises a massive deposit of silty organic sediment that grades upward to a fibrous peat. Accordingly, organic content increases up-core within facies D of core Ok3, from 31% to 90% (Fig. 5). However, this trend is interrupted at 17 cm depth where very fine sand forms a secondary component to the deposit, increasing in concentration to a distinct bed of very fine sand between 5 cm and 12 cm depth that is in sharp contact with the enclosing peat. Organic content decreases to 18% within this sand bed. A photomicrograph shows this to be the finest-grained sand bed within core Ok3. Grains are less than 0.125 mm diameter, moderately sorted, sub-rounded to sub-angular quartz with shell fragments. The age profile for facies D is constrained by four radiocarbon dates derived from organic silt (119 cm depth) and sandy peat (102 cm, 74 cm and 29 cm depth). These yielded conventional radiocarbon ages of 19,942±131 (Wk-21602), 5,117±30 (Wk-25426), 1,467±48 (Wk-21601) and 527±47 (Wk-21600) years BP, respectively (Table 2, Fig. 5).

As noted above, facies Q was only recovered in core Ok1 where it forms the basal unit below 105 cm depth (Fig. 4). This facies is characterised by dark brown fine sand that is dry and compact. The sands are well sorted and massive with no macroscopic organic material observed. The contact with the overlying, softer fine-grained sediments of facies B is sharp and is interpreted as an unconformity based on the different degree of weathering between the two deposits.

5.2. Pollen

All samples from core Ok3 had sufficient pollen for analysis, except the lowermost two at 300 and 284 cm depth. The lower half of the core, incorporating facies A, B and C (below ~120 cm depth) is dominated by Dracophyllum and Myrsine pollen (Fig. 6). Asteraceae, Coprosma and Pseudopanax pollen also feature. Wetland pollen types (in this case Gleichenia and Restionaceae) have low values. Immediately above 120 cm depth, at the base of facies D, pollen and spores of Hebe, Poaceae (grasses) and Pteridium (bracken) peak briefly, and Dracophyllum, Myrsine, Gleichenia and Restionaceae pollen declines. This is coincident with a major increase in Cyperaceae (sedges) pollen followed by an increase in Coprosma and Astelia pollen, Blechnum spores and monolete...
fern spores. Also associated with this is the appearance of *Phormium* pollen. *Pteridium* spores increase sharply at 20 cm depth, in association with increased content of very fine sand within facies D. *Triglochin* pollen and Anthocerotaceae spores appear for the first time at this depth, the former becoming abundant. The long-distance pollen types are from the New Zealand mainland. European-introduced pollen types, namely *Pinus* and *Plantago lanceolata*, were found in the uppermost sample, from the uppermost sand bed (at 9 cm depth). Fragments of microscopic charcoal were present in most of the pollen samples, throughout the full depth of the core.

### 5.3. Foraminifera

Samples from 3 to 4 cm and 7 to 8 cm depth in core Ok3 yielded foraminifera while the other sample at 10–11 cm was barren. The uppermost sample contained two foraminifera specimens that were extremely abraded and therefore only identifiable to suborder Rotaliina. The lower sample is taken from the bed of very fine sand in facies D (5–12 cm) and contained 15 specimens that were also abraded. From these, two species were identifiable; *Ammonia ooteana* (Finlay) and *Elphidium* sp. cf. *crispum*. These two taxa are common in estuarine and inner shelf settings throughout New Zealand (Hayward et al., 1999) and the southwest Pacific (Yassini and Jones, 1995; Strotz, 2003).

### 5.4. Geochemistry

The sampled interval of core Ok1 (0–50 cm) falls within facies D, which extends to 70 cm depth in this core. It comprises organic-rich silt that grades up-core to fibrous peat over the uppermost 30 cm. The deposit also includes two beds of very fine sand; a well defined bed at 43–47 cm and a diffuse deposit at 6–9 cm depth (Fig. 4). Within the sampling interval, organic content ranges from 20% in the organic silt below the deeper sand bed, to 13% in the sand bed, and increases to a maximum of 90% immediately above the sand (Fig. 7). The organic content decreases up-core to about 50% at 15 cm depth and to a minimum of 3% in the diffuse sand bed at 6–9 cm, with 22% organic content in the surface sediments.

The distribution of measured elements through the upper 50 cm of core Ok1 follows two inverse patterns (Fig. 7). Sulphur, chlorine, and bromine, along with the organic matter content, have highest concentration at the base of the peat layer and decrease up-core except for a slight increase in the surface sample (1–2 cm depth). These elements also have low concentrations in the fine sand bed that underlies the peat at 43–47 cm depth. Magnesium follows a similar but less distinct trend (Fig. 7). In contrast, vertical profiles for iron and sodium concentrations are inversely related to the organic content. There is a slight increase up-core through the peat and relatively high concentrations in the fine sand bed (43–47 cm) but with maxima at the base of the sampled interval (48–50 cm). Calcium and strontium concentrations increase from the sand bed to the peat with variations in their up-core profiles, suggesting that they are not solely associated with the mineral matter in the peat.

### 6. Interpretation and discussion

The facies succession in core Ok3 provides the most representative record of wetland evolution and environmental change at Okawa and
Fig. 6. Percentage pollen diagram, Okawa core OK3 (curves have x5 exaggeration in outline).
allows for an evaluation of the impact of tsunamis on this coast. The radiocarbon chronology reveals that the sediment record spans the late Pleistocene to late Holocene, and is a distinctly condensed record (~0.2 m) for the last 500 years. Significantly, the greater part of the sediment record represents terrestrial environments. Thus, facies A is interpreted as a fluvial deposit, on the basis of its coarse-grain size and poor sorting. The single sample (260 cm) that yielded countable pollen is dominated by Asteraceae and Myrsine pollen and monolette fern spores, indicating that local vegetation was woodland, possibly often disturbed. The abrupt transition to the fine-grained sediments of facies B is interpreted to represent a change from an ephemeral stream to a stable, low energy depositional environment that favoured peat accumulation. Pollen in the peat shows Dracophyllum becoming co-dominant with Myrsine in local woodland. These taxa were sub-dominant (~20%) in the pollen record for the tablelands of Chatham Island during the early glacial period (i.e. Oxygen Isotope Stage 3) (Mildenhall, 1994; McGlone, 2002). Radiocarbon ages from Okawa are consistent with this, ranging from ca. 43,000 yrs BP to ca. 37,200 yrs BP (42,200–38,390 cal BP). Low values for wetland and other open vegetation taxa suggest that the woodland at Okawa was dense.

Cool climate conditions are apparent in the pollen record associated with the sand beds of facies C. In particular, at c. 120 cm depth the decline in Dracophyllum pollen coincident with peaks in Asteraceae, Hebe and Poaceae pollen and peak (albeit slight) in Pteridium spores suggest a change in local vegetation from dense woodland to open shrubland. The sand beds at this interval are interpreted as reworked materials washed into the wetland from the sandy soils of adjacent slopes, on the basis of the well sorted, rounded character of the sand grains. The preservation of organic detritus as discrete silt flasers within facies C is a strong evidence for deposition in standing water, but on an episodic basis. The calibrated radiocarbon age of 24,260–23,420 cal BP yrs BP at the contact between facies C and D suggests that these conditions existed immediately prior to the onset of the glacial maximum.

Sedimentation during the glacial maximum appears to have been slow and may not be recorded in the wetland, as suggested by the pollen record at the base of facies D. This record shows clear evidence for a change to a wetter and warmer (post-glacial) environment and increase in peat production. Cyperaceae became the dominant wetland taxon (with Phormium a minor component) in the now wetter substrate, which had increased in area. Coprosma became the dominant woody species on drier ground, with Astelia herbs also common. Blechnum and other ground ferns (monolette) also featured in the local plant community. The timing for this transition to post-glacial conditions at Okawa is not clear. However, calibrated radiocarbon age data from core Ok3 indicate that a peat wetland had developed by 5910–5720 cal yrs BP. This implies the coincident development of a coastal dune system.

Within the upper 20 cm of facies D in core Ok3, both the pollen and sediment record show evidence suggesting disturbance to the dune system at Okawa. Key indicators include the increase in Pteridium and Anthocerotaceae spores and Coriaria pollen at 20 cm depth, microscopic charcoal and the diffuse deposit of very fine sand at 17–12 cm depth (Fig. 5). These changes probably reflect fire disturbance of dune vegetation at Okawa and resultant aeolian transport of dune sand into the wetland, albeit as a small amount. On the New Zealand mainland, the aforementioned pollen and spore types are associated with early Polynesian deforestation by fire. The radiocarbon age from 29 cm depth of 560–470 cal yrs BP provides a maximum age for this disturbance and is consistent with McFadgen’s (1994) suggested date of ca. 450 yrs BP for the earliest human impact on Chatham Island. The coincident appearance of pollen of Triglochin, a herb that favours damp ground, appears also to be related to human activity. Whether the fire disturbance recorded by the charcoal in the diffuse sands (17–12 cm depth) was due directly to burning by people is not known. An alternative explanation is that tsunami inundation of the dunes led to mobilisation of sand into the wetland. This is considered less likely, however, in view of the lack of supporting evidence in this interval of the core. On balance, the diffuse sand is interpreted to represent the initial stages of sand accumulation during the Kekerianean phase at about the same time as humans were first lighting fires in the area.

The diffuse sand bed discussed above is interpreted as a different deposit to the distinct bed of very fine sand that lies above it (5–12 cm depth in core Ok3). One important difference is the trace presence of...
abraded shell and identifiable foraminifera in the sand bed. Based on the known ecological distribution of the two taxa present it is likely that this deposit includes sediment from a shallow water, marine setting. *Ammonia aoteana* is found in a broad suite of ecological conditions, from fully marine through to brackish conditions (Strotz, 2003), whereas *Elphidium crispum* is common in sandy, shallow environments of normal marine salinity (Hayward et al., 1997). The strongly abraded condition of these foraminifera suggests that they were subject to post-mortem transport from their habitual environment, an interpretation consistent with their preservation in a sand bed that includes abraded shell fragments. Together, the texture and composition of the sand bed indicate reworking by wind or water from the local beach-dune system. Of the two transport mechanisms, water is considered more likely, as the deposit forms a discrete bed with sharp contacts and almost one-fifth (18%) of the sediment mass is organic debris that is interpreted as rafted.

Additional proxy evidence to support the interpretation of water as the transport mechanism for the sand bed in facies D is provided by the geochemical analysis of core Ok1. In making this correlation between cores Ok1 and Ok3, the different depths of the sand bed in each core are acknowledged (43–47 cm and 5–12 cm, respectively), however these differences are partly due to different core compaction values (10 cm in Ok1 and 29 cm in Ok3); and partly to topographic differences in the wetland at the time of deposition, being 0.1–0.2 m higher further inland at site Ok3 (Table 1).

In core Ok1, the down-core profiles of several chemical elements indicate the presence of marine waters at the site. In particular, raised values for sulphur, chlorine and bromine, directly above the sand bed at 43–47 cm provide a consistent signature of higher palaeosalinity. These elements have all been used in previous studies as proxies of paleosalinity in coastal deposits because they occur in higher concentrations in seawater than in freshwater (e.g. Minoura et al., 1994; Chen et al., 1997; Chagué-Goff and Goff, 1999; López-Buendia et al., 1999; Chagué-Goff et al., 2000, 2002, Chagué-Goff, 2010). The sulphur peak associated with high organic content suggests that sulphur is present in organic form, which is a common characteristic of clay and peat in brackish environments, as a result of sulphate uptake from seawater (e.g. Lowe and Bustin, 1985; Dominik and Stanley, 1993; Chagué-Goff et al., 2000). Calcium, strontium and magnesium concentrations also increase directly above the sand bed in core Ok1, a rise that can be attributed to contributions from marine micro-organisms carried in the seawater.

Sodium is also often used as a proxy for seawater influence in marsh sediments (López-Buendia et al., 1999), but in core Ok1 it does not record a peak at the same depth as the other elements. A similar result was reported by Chagué-Goff and Fyfe (1996) who found peaks in bromine and chlorine in freshwater peat at a site 10–15 km from the sea, but no associated sodium peak. This may be due to the high solubility of sodium in peat (e.g. Damman, 1978), or to its occurrence mainly in the inorganic fraction of the peat deposit, as indicated by the

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**Fig. 7.** Distribution of chemical elements (S, Cl, Br, Na, Ca, Sr, Mg and Fe) plotted against organic content and sediment facies for the sampled interval (0–50 cm depth) of Okawa core Ok1.

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[Image of Fig. 7: Distribution of chemical elements (S, Cl, Br, Na, Ca, Sr, Mg and Fe) plotted against organic content and sediment facies for the sampled interval (0–50 cm depth) of Okawa core Ok1.]
higher sodium concentrations in the sandy samples below the peat (nearly 1%). Chlorine and bromine on the other hand are transformed to organic forms (e.g. Biester et al., 2004) and therefore retained.

The chemical signature at the base of the peat that overlies the sand bed in core Ok1 suggests a distinct marine influence at the site, but without influx of clastic sediment. The sand bed below provides a record of sediment influx but these sediments do not hold the same geochemical evidence of marine inundation. This apparent contradiction can be attributed to the propensity for sulphur, chlorine and bromine to associate with organic matter (e.g. Lowe and Bustin, 1985; Biester et al., 2004; Chagué-Goff, 2010); hence the marine geochemical signature is missing from the sand bed. In addition, post-depositional processes which may be associated with circulation of groundwater in a sand layer after burial can result in leaching, dissolution and remobilisation of elements and minerals, and have been suggested to explain the absence of saltwater indices in palaotsunami deposits elsewhere (Chagué-Goff, 2010). Similarly, Switzer and Jones (2008) attributed the lack of carbonate microfauna in a tsunami sand sheet to dissolution by organic acids. Active decomposition of peat is known to produce organic acids, and Clarkson et al. (2004) have reported acidic conditions in the peatlands on Chatham Island. Therefore, post-depositional leaching and dissolution of elements and microfauna by acidic groundwater cannot be ruled out.

It is therefore proposed that marine inundation was initiated by an event that drove the transport of beach-dune sands into the wetland at least as far as core site Ok3, a distance of 500 m from the present coastline. Following the initial inundation, marine waters continued to enter the wetland for an unknown time, producing the elevated sulphur, chlorine and bromine concentrations in the peat until the wetland accreted to an elevation above marine inundation, or the dunes closed off the inlet.

This interpretation raises the question of whether the inundation of Okawa wetland was caused by storm surge or tsunami. Given that a deposit of sea-rafterd Loisels pumice (dated to 660–510 cal yrs BP; McFadgen, 1994) exposed in forereads near Okawa Point only reaches as high as modern storm surge deposits, it seems unlikely that any storm surge has ever risen high enough to enter Okawa wetland, at least with the 8 m high dunes in place. Tsunami inundation of the wetland is more plausible, an interpretation that is supported by the anecdotal record of ~7 m run-up height for the 1868 tsunami. That the actual age of the putative tsunami sand deposit in Okawa wetland can be tied to the 1868 event is given indirect support by the pollen record in core Ok3, which shows that pollen from European exotics (Pinus and Plantago lanceolata) is preserved within the tsunami sand. Clearly, this indicates that the inundation occurred during historical times, with the 1868 event the most likely candidate. The cumulative proxy evidence from cores suggests that the tsunami inundated Okawa wetland, via the stream channel, and deposited a thin sheet of fine shelly sand in the process. The same tsunami most likely deposited beach gravels and pebbles on the Okawa dunes.

7. Conclusion

The 1868 tsunami was a major Pacific-wide event that left a subtle but tangible sedimentary signature on Chatham Island. That the depositional record in Okawa wetland was not more extensive, or incorporate sediments of a greater calibre (e.g. gravel), can be attributed to attenuation of flow velocity as the tsunami crossed the dunes. Reworking of the dunes was minimal, however, as evidenced by preservation of dune palaeosols and the Loisels pumice deposit (McFadgen, 1994), that pre-date the 1868 tsunami. Rather, it appears that the Okawa dunes acted as a very effective buffer to the tsunami surge, forcing it to run-up the dune face, as implied by the anecdotal record of the event (de Lange and McSaveney, 2009).

The pollen and radiocarbon evidence from Okawa wetland show that the sediment record is largely a legacy of late Pleistocene environments, with the Holocene phase condensed to the upper metre. This in contrast to other coasts where multiple palaotsunami events spanning the mid to late Holocene have been reconstructed from sediment records that are typically several metres thick (e.g. Cisternas et al., 2005; Peters et al., 2007; Goff et al., 2010b). Okawa wetland therefore represents an end-member setting on the spectrum of accommodation space that is opposite to those wetlands that have experienced co-seismic subsidence in association with tsunami inundation (e.g. Atwater, 1987). Clearly, the latter situation favours preservation of multiple tsunami events over geological time, whereas the lack of accommodation space at Okawa does not (Goff et al., 2008). For Chatham Island, Okawa wetland is not likely to be a unique case, as most of the coast is fronted by high dunes and other swamps and wetlands are expected to have similarly condensed Holocene sediment records (McGlone, 2002). Thus, while the datasets for palaotsunami events in the Pacific Ocean are growing (e.g. Peters et al., 2003; Goff et al., 2010a), future efforts to extend and correlate these records at regional or ocean basin scales, requires careful selection of sites where sediment records are not condensed and tsunami inundation is likely to have been unrestricted.

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References

Anonymous, 1868. The Earthquake Wave at the Chatham. Lyttelton Times, August 31 1868.