



Lagoon subsidence and tsunami on the West Coast of New Zealand

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Abstract

Sediment core and trench data from a coastal lagoon on the West Coast of the South Island, New Zealand are used to investigate evidence for co-seismic subsidence and associated tsunami inundation. Physical data are used to document a salt marsh soil buried ~80 cm below the modern sediment surface that is locally covered by a gravelly sand bed. The sediment record also contains geochemical and biological (diatom and foram) evidence for abrupt changes in salinity of lagoon waters that link to subsidence, tsunami flooding and to the open versus closed state of the lagoon tidal entrance. At the local scale, these relationships allow for separation of tsunami evidence from other agents of environmental change in the lagoon. We also propose a conceptual connection between these local changes and regional drivers of landscape development, most notably major earthquakes and resultant pulses in sediment supply to the coast.

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1. Introduction

Multiproxy analysis of coastal wetland sediments in the Northern Hemisphere has produced important records of past environmental changes, including extreme events such as earthquakes and tsunamis (e.g., Atwater and Moore, 1992; Shennan et al., 1996). Research has generally focused on sites where a demonstrable linkage

can be made between tsunamis and their generating mechanism (e.g., Pacific northwest, Atwater, 1987; Clague, 1997; Clague et al., 2000). This work and work on other palaeotsunami deposits elsewhere (Dawson et al., 1988; Bryant et al., 1992; Minoura et al., 1996) have served to considerably improve our geological understanding of palaeoseismic-related signatures preserved in the sedimentary record. In New Zealand, we have the opportunity to test and develop this approach by utilising a largely undisturbed and unstudied coastal sediment record that has formed in a seismically active setting (Goff et al., 2001).

New Zealand sits astride the boundary between two major tectonic plates, the Australian and the Pacific. In historical times seismic activity has ranged from barely detectable earthquakes and tsunamis that have caused no

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46 damage, to large-scale earth and sea movements (Eiby,
47 1982; Hull, 1986). Recent reviews of past tsunamis in
48 New Zealand estimate that large, locally generated
49 events, possibly with a nationwide impact throughout
50 the North and South Islands, occur about once every
51 500 years (Goff and McFadgen, 2002). The most recent
52 events in New Zealand occurred around the mid 15th
53 century AD and are associated with a cluster of large
54 earthquakes, one of which was an Alpine fault rupture
55 (Goff and McFadgen, 2002).

56 The Alpine fault is over 400 km long, extending
57 almost the entire length of the South Island (Berryman
58 et al., 1992; Bull, 1996; Norris and Cooper, 1997). This

single right-lateral, oblique slip fault raises the Southern
Alps at a rate of about 5–8 m per 1000 years (Bull,
1996) (Fig. 1). The last three major ruptures (magni-
tudes of ca. 8.0 M_w) are believed to have occurred in AD
1717 (along a 375 km segment), ca. AD 1630 and ca.
AD 1460 (involving a minimum 300 km segment, with
a vertical offset of 2.15 ± 0.4 m) (Yetton et al., 1998;
Wells et al., 1999). A smaller rupture occurred in AD
1826 (Wells et al., 2001; Cullen et al., 2003).

This study examines the recent sedimentary record of
Okarito Lagoon on the West Coast of the South Island,
with the aim of testing for evidence of co-seismic sub-
sidence of the lagoon floor and tsunami inundation,

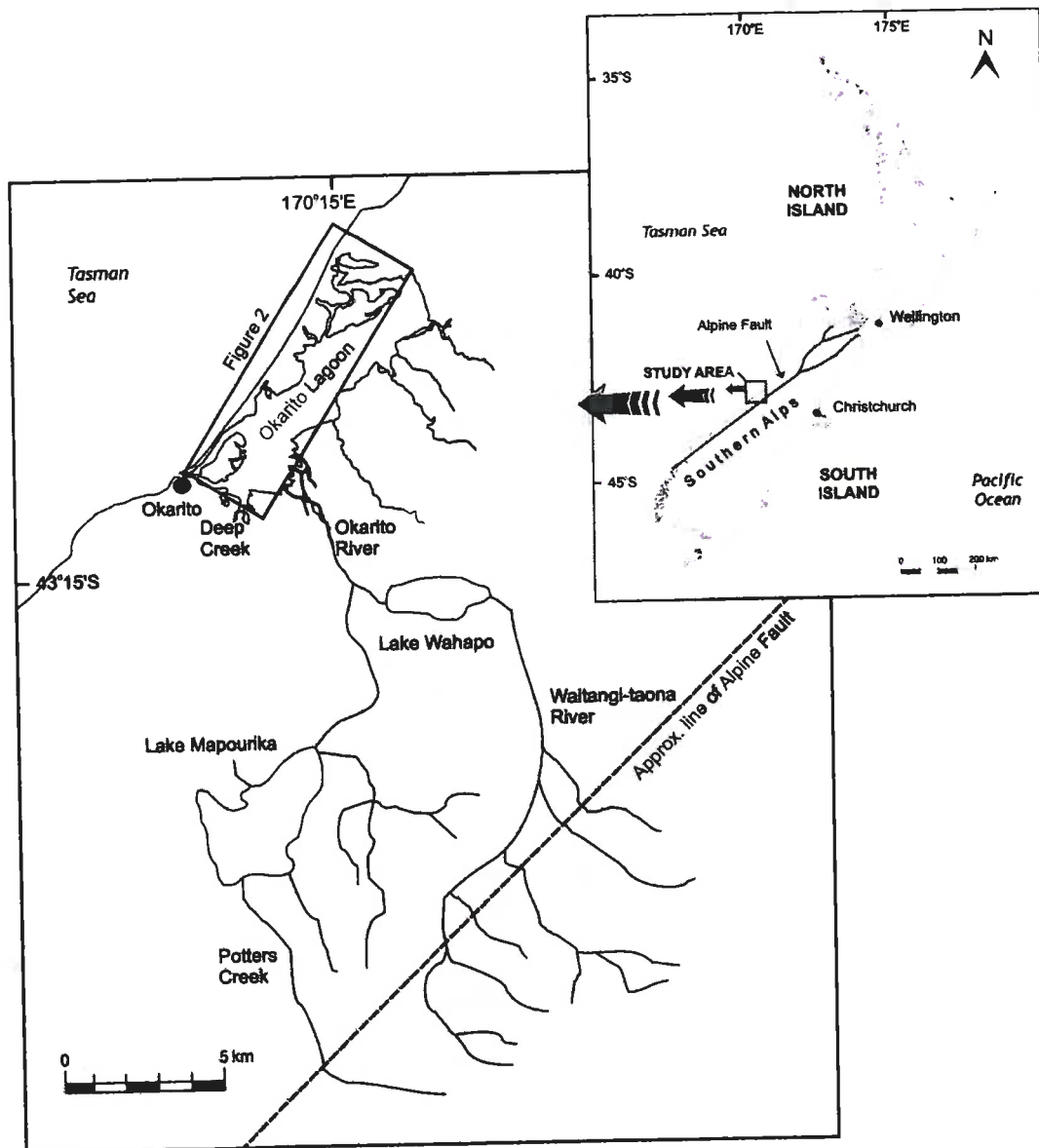


Fig. 1. Location map showing Okarito Lagoon, West Coast of South Island, New Zealand.

72 given the context of frequent Late Holocene seismicity.
 73 To achieve this we undertook a multiproxy analysis of
 74 sediment cores and trenches from the lagoon. This paper
 75 is an extension to earlier work on the same sample set
 76 published by Goff et al. (2004).

77 2. Okarito Lagoon

78 Okarito Lagoon, Westland (43°11'S, 170°14'E) is
 79 one of the largest estuarine inlets on the West Coast of
 80 the South Island. It is 10 km long and 2–4 km wide with
 81 an open water area of ~20 km² and fringing salt marsh
 82 area of ~10 km² (Fig. 1). The lagoon is separated from
 83 the sea by a narrow sand barrier with a maximum
 84 elevation of ~4 m. The width of the barrier decreases
 85 southward from 500 m to 100 m at Okarito village,
 86 where an ephemeral tidal inlet provides the only direct
 87 connection to the open coast. From vertical aerial
 88 photographs it is evident that the lagoon has a complex
 89 history of barrier breaching, with four previous tidal
 90 inlets ending as blind channels against the landward side
 91 of the barrier (Fig. 2). The form of these channels has
 92 been maintained as surrounding marshes have devel-
 93 oped upon probable relict flood tidal delta surfaces. Our
 94 investigation focused on the southernmost marsh island
 95 around which the active inlet channel bifurcates, and on
 96 the tidal flat to the south of the main channel. Tidal
 97 range on the coast varies from 1 m on neaps to 2.1 m on
 98 spring tides. Holocene sea-level history for the West
 99 Coast is poorly understood in detail, so we assume a
 100 similar history for elsewhere in New Zealand which
 101 records stillstand since ca. 6.5 ka (Gibb, 1986).

102 Rainfall in the region exceeds 10,000 mm/year,
 103 resulting in high erosion rates (5–12 mm/year) of the
 104 fractured, easily eroded schists and glacially derived
 105 materials found to the east of the Alpine fault (Berry-
 106 man et al., 1992; Norris and Cooper, 1997). To the west
 107 of the Alpine fault, the overprint of Late Quaternary
 108 glacial activity is evident in the large moraines that
 109 surround Okarito Lagoon (Griffiths and McSaveney,
 110 1986). Freshwater flow into the lagoon is derived from
 111 three sub-catchments; Okarito Forest (9700 ha) which
 112 supplies six small creeks that drain the moraines and
 113 enter along the northeast shoreline, and; Lake Mapourika
 114 (ca. 9000 ha) and Waitangi-taona River (8300 ha)
 115 catchments which drain to Okarito River and Deep
 116 Creek at the southern end of the lagoon. These catch-
 117 ments extend further inland to drain terrain formed in
 118 schist and glacial outwash and are traversed by the
 119 Alpine fault in their upper reaches (Fig. 1). The major-
 120 ity of the sediment load is captured in Lake Mapourika
 121 and Lake Wahapo, however, and it is the moraines that



Fig. 2. Aerial photograph mosaic (unrectified) for Okarito Lagoon taken January 1988, showing core and trench sites for this study and interpreted former tidal inlets (marked as blind channels). Source: Land Information New Zealand, courtesy Department of Conservation. Crown Copyright Reserved.

122 provide the main source of coarse material entering
 123 Okarito Lagoon (MacPherson, 1981). At its mouth,
 124 Okarito River is constructing a small (~1 km²) delta
 125 that has infilled the southeast corner of the lagoon. A
 126 raised shoreline bench, estimated at 1 m above present
 127 water level, fringes the landward side of Okarito La-
 128 goon and is backed by shallow (<5 m) caves cut into
 129 cliffed sections of moraines.

130 An additional key component to landscape develop-
 131 ment in the West Coast region is seismicity. In parti-
 132 cular, ruptures on the Alpine fault have been shown to
 133 have resulted in widespread forest destruction due to
 134 ground shaking (Wells et al., 1998, 2001; Cullen et al.,
 135 2003), river aggradation (Yetton et al., 1998), and rapid
 136 coastal dune building (Goff and McFadgen, 2002); the
 137 latter two being direct consequences of increased sedi-
 138 ment yield from hillslopes following earthquake-
 139 induced landslides. It is in this context of strong seismic
 140 overprinting on the coastal landscape that we recognized
 141 the potential for a record of seismic events to be re-
 142 corded in Okarito Lagoon.

143 3. Data sources and analytical techniques

144 Sediment cores and trenches were used to investigate
 145 the shallow stratigraphy of southern Okarito Lagoon.
 146 Cores were collected by the vibrocore technique at two
 147 sites from the lagoon–marsh interface and close to high
 148 water mark (Fig. 2). Compaction was measured prior to
 149 core recovery. In the laboratory, cores were split
 150 lengthwise, logged and sub-sampled for grain size,
 151 organic content, micro- and macrofossils, geochemical,
 152 and radiocarbon analyses. Trenches (1–3) were dug by
 153 hand to confirm the lateral continuity and character of
 154 sedimentary units observed in the cores. Additional
 155 sediment samples were collected from trenches. Grain
 156 size was measured on 23 core and 12 trench sub-samples
 157 using a laser particle sizer (Galai™) system that
 158 determines particle size using the time-of-transition
 159 principle (Molinarioli et al., 2000). Results are reported
 160 for particle volume measurement. For trench 1, we
 161 determined grain size on two samples by mechanical
 162 sieving at half-phi intervals from –1 to 4 phi. Organic
 163 content of cores was determined via loss-on-ignition
 164 (LOI) treatment of 17 sub-samples by ashing at 500 °C for
 165 4 h. Results are reported on a dry weight basis. Elemental
 166 analysis of seven sub-samples from core 6 was undertaken
 167 using ICP-AES, following sample preparation by a multi-
 168 acid ‘total’ digestion. Geochemical data have been
 169 normalised for grain size following Loring (1991).

170 For diatom analysis, 15 sub-samples from core 6
 171 were prepared following standard techniques (e.g., Bat-

172 tarbee, 1986). Fossil diatoms were mounted in Naphrax
 173 and counted using light microscopy at a magnification
 174 of ×1000. A minimum count sum of 600 diatom valves
 175 was used. Identification of diatom taxa and their palaeo-
 176 environmental interpretation is based upon established
 177 floras and other texts, including Hustedt (1927–1966,
 178 1930, 1957), Van der Werff and Huls (1957–1974),
 179 Cleve-Euler (1951–1955), Krammer and Lange-Bertal-
 180 lot (1986–1991), Hendey (1964), Round et al. (1990),
 181 Simonsen (1967), De Wolf (1982), Admiraal (1984),
 182 Denys (1991a,b), Vos and de Wolf (1993a,b, 1994),
 183 Foged (1977, 1978, 1979), John (1983), Crosby and
 184 Wood (1958, 1959), Wood et al. (1959), and Wood
 185 (1961, 1963). Results are presented as a percentage
 186 frequency diagram, showing the key diatom species
 187 within salinity (Halobion) groupings.

188 For foraminiferal analysis, eight 10-cm³ samples of
 189 sediment were processed from cores 4 and 6. Samples
 190 were washed over a 63-µm sieve to remove mud and
 191 heavy liquid floatation used to concentrate foram tests in
 192 sandy samples. Census counts of all benthic forams
 193 were made for each sample. Estimates of the tidal ele-
 194 vation at which each foram species was deposited were
 195 made using the Modern Analogue Technique (Hayward
 196 et al., 2004), based on relative abundance data for 250
 197 benthic foram fauna from modern estuaries and coastal
 198 lagoons around New Zealand (data from Hayward et al.,
 199 1999a). Similar techniques to determine estimates of
 200 tidal elevation based upon foram assemblages have been
 201 used elsewhere (Scott and Medioli, 1986; Van de Plas-
 202 sche et al., 1998). In this study, elevation estimates are
 203 presented as tidal range derived from the five most
 204 similar modern faunas in the analogue set. The reli-
 205 ability of these estimates depends on a number of
 206 factors, including the range of tidal levels; depths and
 207 environments represented by the analogue samples; and
 208 the breadth of the tidal and depth ranges of the major
 209 taxa. Previous studies (e.g., Hayward et al., 1999b) have
 210 shown that the most precise tidal ranges can be obtained
 211 near high tide level from marsh faunas, with far broader
 212 ranges observed in intertidal mud and sand flats and in
 213 subtidal environments. Some taxa with restricted high
 214 tidal ranges in near normal salinity situations (e.g.,
 215 *Haplophragmoides wilberti*, *Trochammina salsa*, *Mili-*
 216 *ammina fusca*) are known to live in abundance through-
 217 the entire tidal range and also subtidally in more brack-
 218 ish environments (e.g., Hayward et al., 1999a). Thus an
 219 assessment of the setting and probable salinity of the
 220 lagoon or estuary at the time is important in estimating
 221 tidal level.

222 Foram data also provided for calculation of an arti-
 223 ficial salinity index (SI). This was done using detrended

224 and canonical correspondence analysis, based on the
 225 relative abundance of foram species following Hayward
 226 et al. (2004). SI estimates range between 0 (freshwater)
 227 and 10 (normal marine salinity).

228 4. Results

229 4.1. Stratigraphy and sediment texture

230 All five sample sites preserve a similar stratigraphic
 231 record within the upper metre (Figs. 3–5). The
 232 bulk of the sediment matrix comprises massive beds

233 of fine to medium sand and silt with local shell and
 234 wood fragments. Within this, each core and trench
 235 recovered a buried marsh soil at variable depths,
 236 ranging from 50 cm (trench 3) to 88 cm (trench 1)
 237 with depths in the cores falling within this range. The
 238 soil unit is olive-brown to grey-olive silt that encloses
 239 fine root hairs and wood fragments. It is in sharp
 240 upper contact with overlying sediments that are dis-
 241 tinguished by a colour change to light grey and ab-
 242 sence of organic material, an observation supported
 243 by LOI results which show a 1–2% decrease across
 244 this contact (Fig. 6).

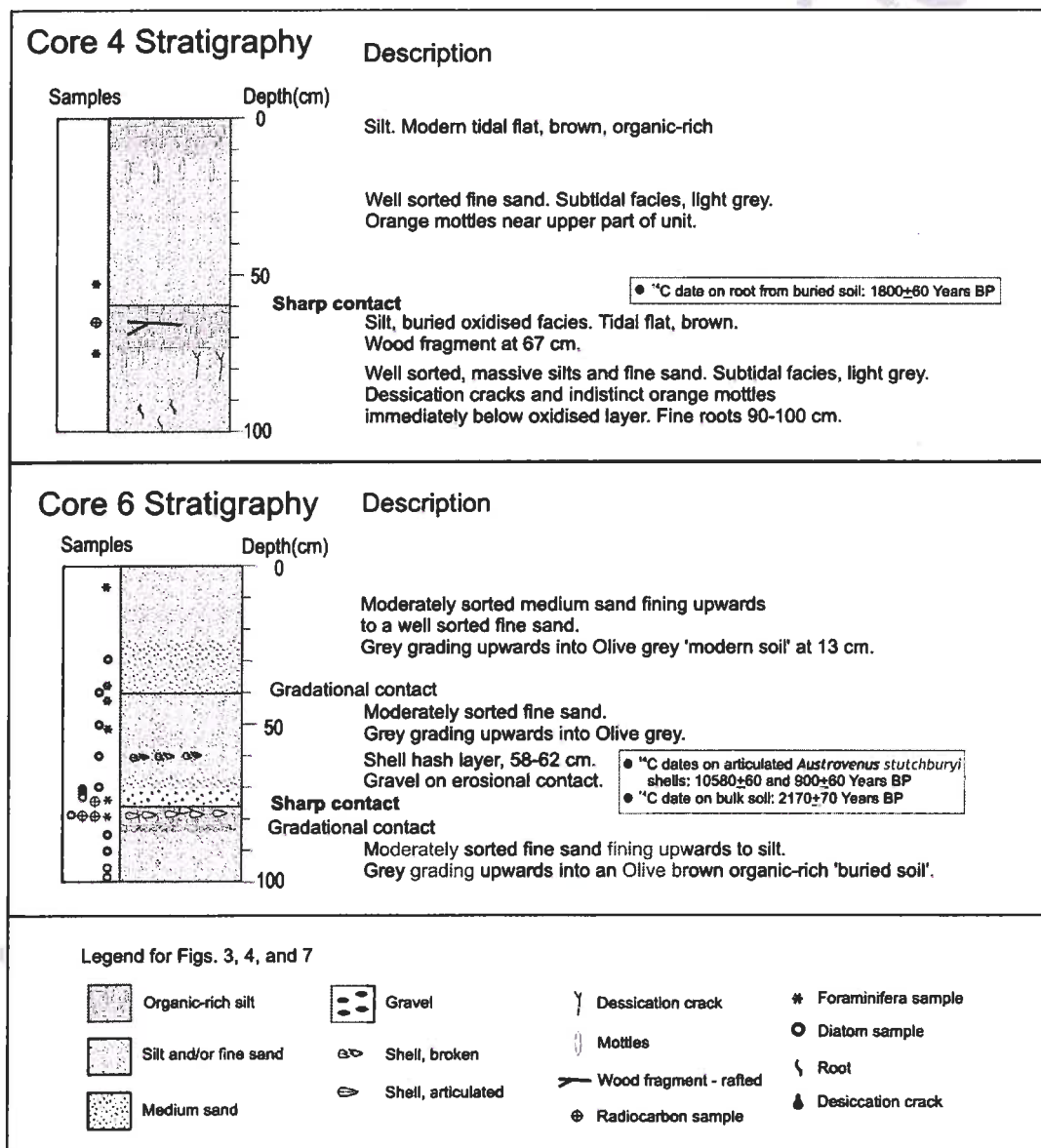


Fig. 3. Graphic logs for cores 4 and 6.

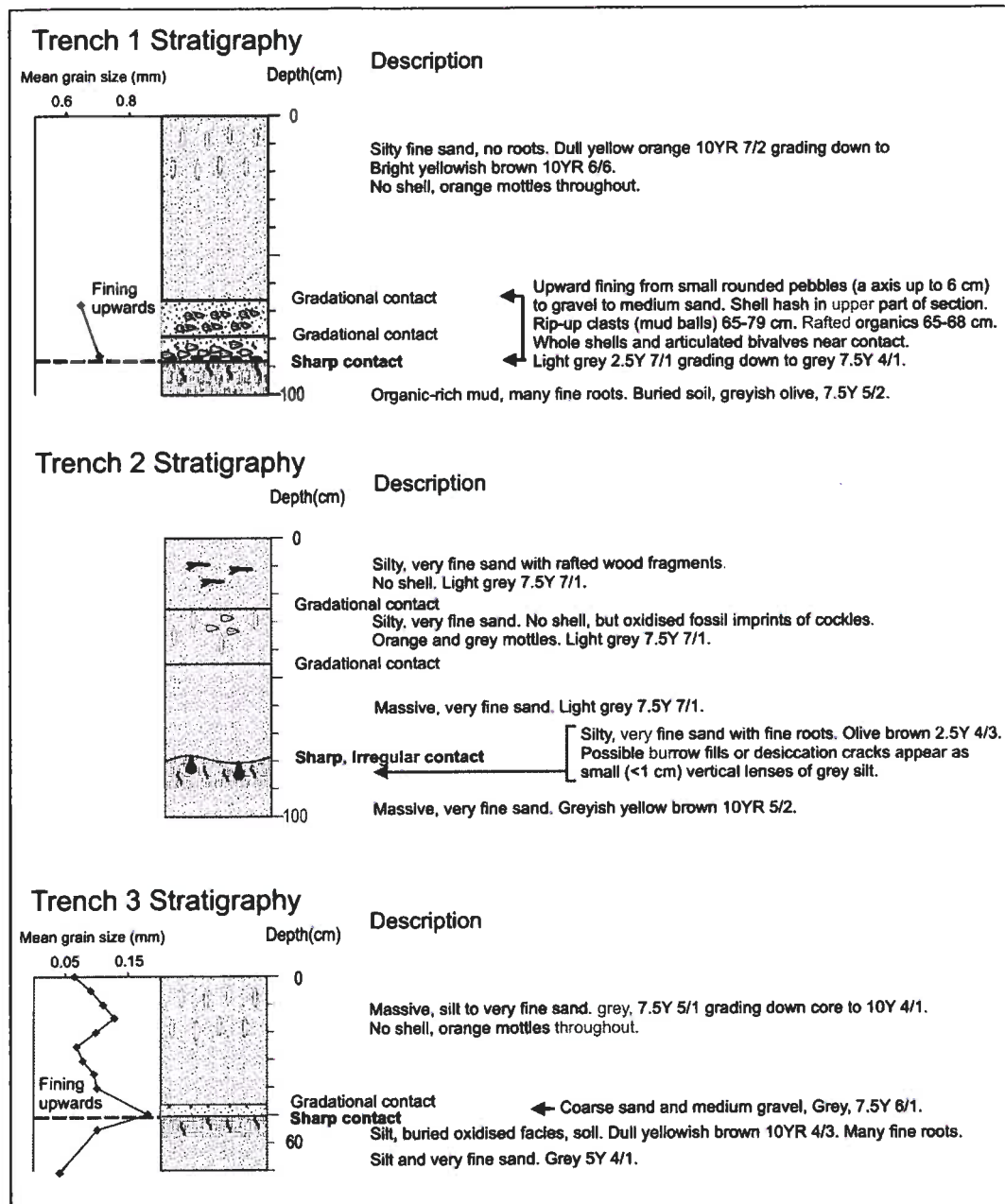


Fig. 4. Graphic logs for trenches 1–3.

245 In trench 1, the buried soil is overlain by a 23-cm-
 246 thick bed of shell-bearing gravel and coarse sand. Gra-
 247 vels are rounded and shell remains include whole and
 248 articulated bivalves near the base of the unit that
 249 become more commonly broken toward the upper con-
 250 tact at 65 cm depth. Shell species are dominated by the
 251 cockle *Austrovenus stutchburyi*, with rare examples of
 252 the intertidal bivalve *Macra ovata* and the gastropod
 253 *Amphibola crenata*. In addition, remains of macro

254 algae, most likely seaweed, were recovered. On the
 255 landward side of the lagoon (core 4 and trench 2)
 256 sediment above the soil comprises 20–30-cm-thick beds
 257 of very fine sand that are massive and carry shell imprints
 258 and wood fragments. In trench 2, the contact is on an
 259 irregular surface and grey fine sands have filled vertical
 260 cracks that penetrate 10 cm into the buried soil; these
 261 cracks may have originated as burrows or, more likely,
 262 via desiccation of the soil when exposed subaerially. In

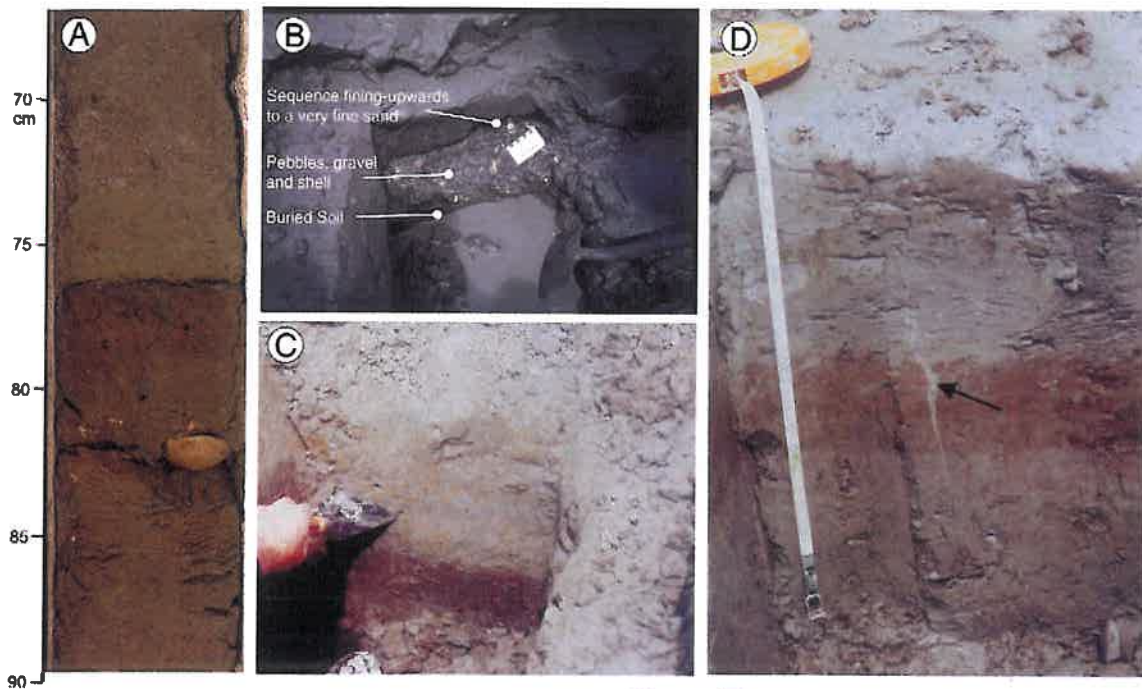


Fig. 5. (A) Core 6 showing soil surface buried 77 cm below modern tidal flat and articulated bivalve (*Austrovenus stutchburyi*) dated to 900 ± 60 years BP; (B) trench 1 showing buried soil overlain by rounded pebbles, gravel and shell, grading upward to very fine sand. Scale card is 10 cm long; (C) trench 2 showing buried soil overlain by very fine sand; arrow indicates infilled desiccation crack or burrow. Tape is 50 cm long; (D) trench 3 showing buried soil overlain by fine sand.

263 trench 3, the buried soil is overlain by a 4-cm-thick bed
264 of coarse sand and medium-sized gravel that grades
265 upward to a 40-cm-thick deposit of very fine sand.

266 4.2. Geochemistry

267 A suite of elements assayed from core 6 show a clear
268 peak in concentration within the silty sands that overlie
269 the buried soil (Fig. 6). Thus, iron (Fe), sulphur (S),
270 titanium (Ti), strontium (Sr), barium (Ba), calcium (Ca)
271 and sodium (Na) all increase markedly at 77 cm depth
272 and then decrease within the successive bed of fine-
273 medium sand to 20 cm depth. Although S occurrence in
274 sediments is mostly related to post-depositional diage-
275 netic processes, Fe-rich sediments are more likely to fix
276 S than Fe-poor sediments. Thus S and Fe can be used
277 together as indicators of palaeosalinity in sediments
278 (Thomas and Varekamp, 1991; Daoust et al., 1996;
279 Chagué-Goff and Goff, 1999). Na has also previously
280 been used as a proxy for marine influence in marsh and
281 lagoon systems, because of the relatively high Na con-
282 tent in seawater compared with freshwater (López-
283 Buendía et al., 1999; Chagué-Goff et al., 2002). Simi-
284 larly, Sr, Br and Ca also occur in higher concentrations
285 in marsh and wetland sediments inundated by saltwater

(Minoura et al., 1994; Chen et al., 1997). Titanium is
used here as an indicator of heavy mineral concentration
(e.g., magnetite). Therefore, the peak Ti concentration
directly above the buried soil within core 6 is interpreted
to record the preferential sorting and concentration of
heavy minerals. In a lagoon, this particular response
would most likely occur during a high energy event such
as initiated by tsunami, storm surge or large river flood.
Collectively, the measured elements all indicate a sud-
den rise in the salinity of lagoon waters, followed by a
gradual return to brackish conditions.

4.3. Diatoms

Diatom analysis of core 6 provides further evidence
of palaeoenvironmental change for Okarito Lagoon
(Fig. 7). Fossil diatoms are generally well preserved and
abundant, providing a flora of >240 taxa. In the lower
part of the core (87–100 cm depth) brackish water taxa
(mesohalobion, M; oligohalobion, OI-indifferent and
OH-halophile) comprise >50% of the diatom assem-
blage. *Melosira juergensii* and *Mastogloia elliptica* do-
minate the M taxa, with lower but significant values of
Mastogloia pseudoexigua, *Cyclotella operculata* var.
mesoleia and *Rhopodia gibberula* (OH — indicative

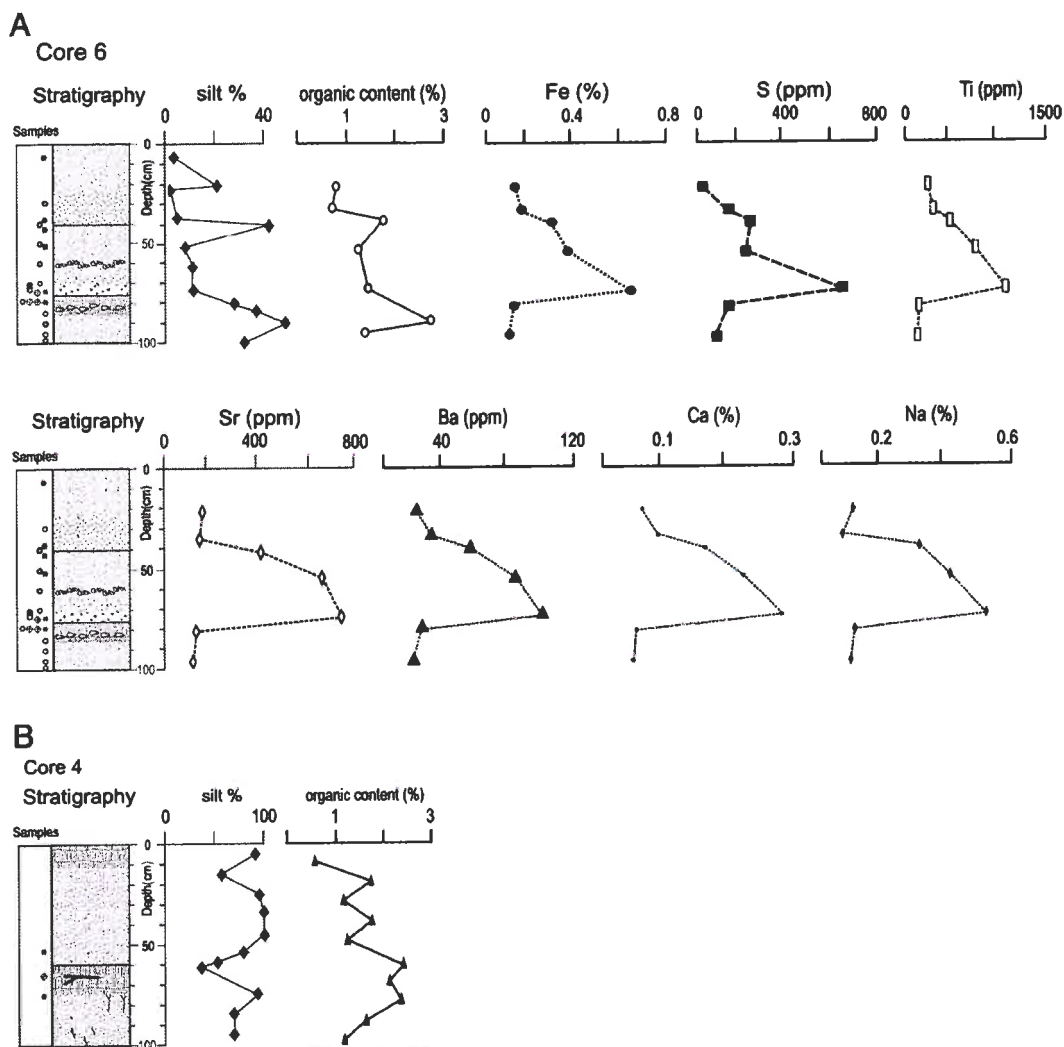


Fig. 6. (A) Geochemical data for core 6, also showing trends in silt and organic content. (B) Trends in silt and organic content for core 4.

309 of sandy conditions). The OI group, rich in *Cocconeis*
 310 *placentula* and *Cyclotella stelligera*, maintains frequen-
 311 cies >32% throughout the interval. There is a distinct
 312 background marine influence (polyhalobion, P), but at
 313 frequencies <15%. Marine diatoms are represented
 314 by *Diploneis suborbicularis*, *Raphoneis surirella*,
 315 *Raphoneis minutissima*, *Thalassiosira eccentrica* and
 316 *Paralia sulcata*. These are littoral species, as are many
 317 of the M group; together representing shallow marine-
 318 brackish water of sandy foreshore, sand flat and mud
 319 flat environments (Denys, 1985). Open coast marine
 320 species are also recorded in the sediments, confirming
 321 tidal exchange at the site. These marine diatoms are
 322 sparse and often broken, indicating their presence is
 323 probably the result of progressive inwashing and re-
 324 working of more open water, sandy marine sediments.

The interval between 67 and 87 cm depth is dis- 325
 326 tinguished by alternating peaks of marine (P) and
 327 brackish–freshwater (OH, OI) groups. Marine diatoms
 328 have peak concentration at 85 cm (20%) and 73 cm
 329 (27%), the former associated with the buried soil and
 330 the latter with the overlying fine sand. Species typical
 331 of sandy–mud flat environments are prominent in
 332 these apparently marine-enriched sediments, with in-
 333 creased values of *Raphoneis* spp., *Cocconeis scutellum*,
 334 *Nitzschia granulata*, *N. punctata*, *Striatella unipunctata*
 335 and *Licmophora* spp. Many other marine species
 336 occur in low numbers, including *Cerataulus turgidus*,
 337 *Actinocyclus octonarius* var. *crassus*, *Grammatophora*
 338 *oceanica*, *Triceratium* spp. and *Navicula lyra*.

A decline in marine taxa (P) occurs above 85 cm 339
 340 depth through the buried soil and high values of brackish

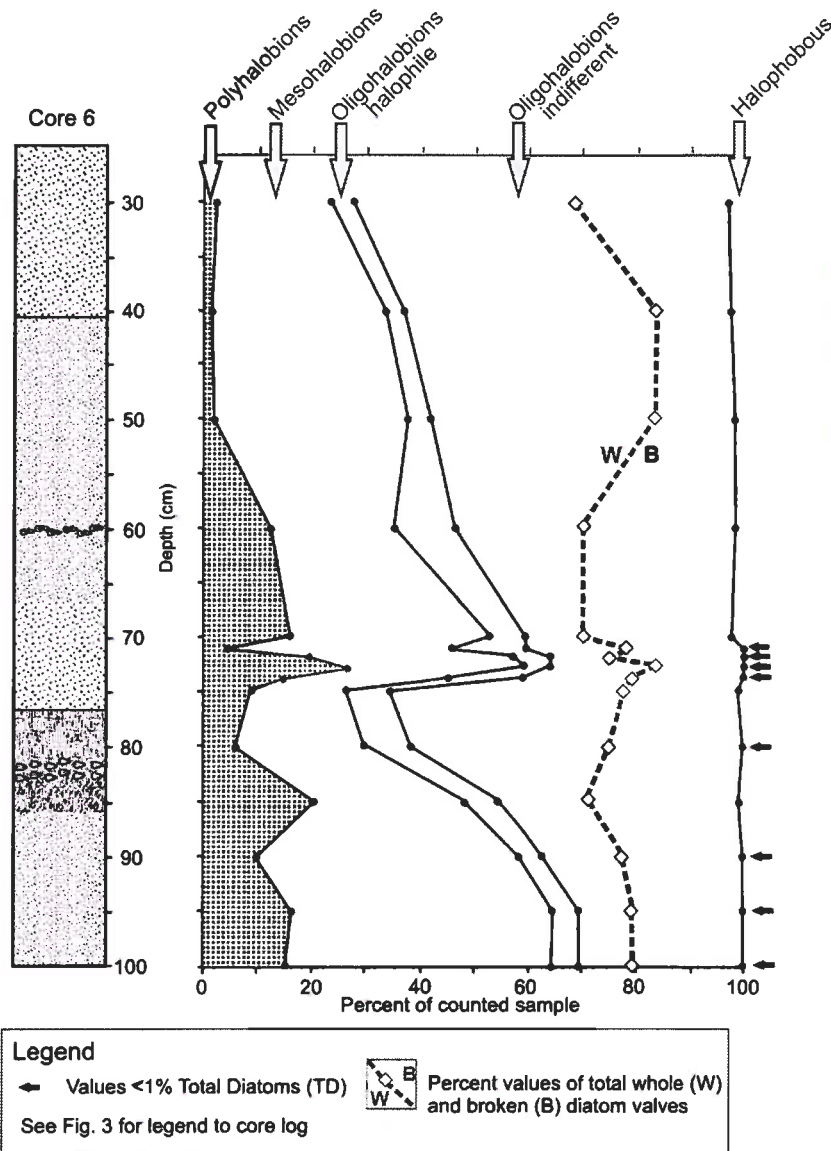


Fig. 7. Summary plot of diatom salinity (halobion) groups for core 6. Polyhalobion (P) — marine conditions, mesohalobion (M) — full brackish water conditions, oligohalobion (OI) — freshwater, salt indifferent; oligohalobion (OH) — some brackish water tolerance; halophobous (H) — freshwater conditions.

341 (OH) and freshwater taxa (OI) are recorded, with indi- 353
 342 vidual maxima for *R. gibberula*, *C. placentula* and 354
 343 *Navicula tripunctata*. Full brackish water species (M) 355
 344 increase above the soil with *Mastogloia* spp., *Melosira* 356
 345 *jeurgensii* reaching peak frequencies at 75–73 cm, with 357
 346 lower values of *Synedra fasciculata* and *Nitzschia* 358
 347 *sigma*. Many of these OI–M group diatoms are well 359
 348 adapted to intertidal–salt marsh conditions and are 360
 349 tolerant of drying–air exposure. 361

350 Between 67 and 30 cm depth, freshwater (OI) species 362
 351 dominate and frequencies of brackish (M) diatoms 363
 352 decline from >30% to ~20%. Above 67 cm the marine 364

signal collapses to <2% above 50 cm. The OI and OH 353
 diatoms (together >60% at the base of this interval) are 354
 joined by a range of other fresh–brackish water species, 355
 increasing to near 80% at 30 cm. These include many 356
Achnanthes spp., with *Achnanthes hungarica* replacing 357
C. placentula as the dominant taxon at the top of the 358
 zone. The diatom assemblage suggests an overall prog- 359
 ressively shallowing, fresh–brackish water and river 360
 flood-dominated environment. 361

Many of the fresh and brackish water diatom valves 362
 (OI/OH and M groups) are relatively unworn and whole 363
 valves dominate the general diatom assemblages (Fig. 7). 364

365 This lack of breakage is unusual given the potential for
366 high-energy conditions and the reworking of sediments
367 at the site. Many of the OI–M diatoms however represent
368 low to medium salinity environments (Cl <20 ppt) and
369 are predominantly of local to *in situ* origin, with growth
370 under intertidal–subaerial conditions. Peaks in key fresh-
371 water (OI) species between 87 and 30 cm are associated
372 with terrestrial plant debris most probably introduced
373 during river flooding. Overall, the diatom record from
374 core 6 shows deposition generally in brackish to brack-
375 ish–freshwater environments of estuary–lagoonal–
376 marsh settings. Toward the top of the sedimentary
377 succession a marked decline in the percentages of marine
378 species and the associated increase in freshwater species
379 indicate an increasing isolation from the sea.

380 The higher resolution diatom analysis between 67
381 and 87 cm shows at least one distinctive marine ‘event’
382 in the sedimentary sequence. It is worth noting that, if a
383 lower sampling resolution had been maintained, then the
384 extremes in diatom values (75–70 cm) would not have
385 been apparent. Two peaks in marine diatoms are re-
386 corded. The first, at 85 cm, coincides with the base of
387 the buried soil. In the context of a consistent marine
388 diatom signal this peak is most likely an artifact of the
389 stratigraphy and sedimentary processes. Marine diatoms
390 tend to be more heavily silicified and resistant to de-
391 struction. Local reworking of sediments to concentrate
392 the marine diatoms present may be the explanation, as
393 supported by the increase in broken valves at this level.
394 The second marine peak (at 73 cm) is more distinctive
395 and associated with the sand bed overlying the buried

soil. This peak is bracketed by higher concentrations of
fresh–brackish water diatoms (e.g., 80 cm within the
soil, 71 cm in fine sand) that we interpret to represent the
prevailing background shallow water, intertidal to salt
marsh environment at the site.

4.4. Foraminifera

Forams extracted from cores 4 and 6 are used here to
reconstruct palaeo-elevation ranges for depositional
surfaces (Table 1). A total of 10 taxa are reported, in
varying abundance and diversity through the sediment
record. In core 6, a sample taken directly below the
contact between the buried soil and overlying sand (79–
81 cm) is co-dominated by agglutinated faunas that
indicate weakly brackish conditions (*Ammotium fragile*,
H. wilberti) on a lagoon or marsh surface that was close to
mean high tide level (range 2–2.2 m above MSL). Within
the overlying sand (71–73 cm), the fauna is strongly
dominated by calcareous taxa (i.e., *Ammonia aoteana*)
indicative of increased salinity and a broader elevation
range spanning subtidal (2.5 m below MSL) to mid-tide
(1.75 m above MSL). Four samples taken from the
interval 64–5 cm in core 6 all indicate a return to lower
salinity, based on consistently high values for *H. wilberti*
and THE strong presence of *A. fragile* and *M. fusca*.
Estimated elevations range from mid-tide (1.9 m above
MSL) to mean spring high tide level (2.2 m above MSL),
most likely in a brackish marsh environment.

Foram concentrations from the two samples taken
from core 4 are very low, allowing estimation of tidal

t1.1 Table 1
t1.2 Foraminifera data for Okarito cores 4 and 6

t1.3		Core 6 ^a					Core 4 ^b			
t1.4	Depth (cm)	5–7	50–52	57–59	62–64	71–73	79–81	101–103	54–56	76–78
t1.5	Number of species	5	3	2	5	6	5	5	1	4
t1.6	Total foram count	104	201	3	25	163	360	250	7	11
t1.7	Relative abundance %									
t1.8	<i>Ammobaculites exiguus</i>	0	0	0	0	0	0	0	0	46
t1.9	<i>Ammonia aoteana</i>	0	0	0	20	77	1	35	0	0
t1.10	<i>Ammotium fragile</i>	13	54	0	4	5	34	19	0	18
t1.11	<i>Elphidium excavatum</i> f. <i>clavatum</i>	0	0	0	0	1	0	4	0	0
t1.12	<i>E. excavatum</i> f. <i>excavatum</i>	0	0	0	0	1	0	0	0	0
t1.13	<i>Haplophragmoides wilberti</i>	64	45	67	68	14	30	34	0	18
t1.14	<i>Jadammina macrescens</i>	1	0	0	0	0	0	0	0	0
t1.15	<i>Miliammina fusca</i>	21	1	33	4	2	14	8	100	18
t1.16	<i>Pseudothurammina limnetis</i>	0	0	0	0	0	21	0	0	0
t1.17	<i>Trochammina inflata</i>	1	0	0	4	0	0	0	0	0
t1.18	Elevation (m above MSL)	1.9–2.2	2.0–2.2	n/a ^c	2.2	–2.5–1.75	2.0–2.2	–2.5–1.15	n/a	–2.5–1.75
t1.19	Salinity index	2.6	2.4	2.6	3.3	5.5	2.2	6.2	3.2	4.3

t1.20 ^a Core 6 from edge of *Leptocarpus similis* salt marsh at neap high tide level, ~1.8 m above MSL.

t1.21 ^b Core 4 from *Leptocarpus similis* salt marsh at mean high tide level, ~2 m above MSL.

t1.22 ^c n/a indicates insufficient foram preservation to allow application of MAT.

t2.1 Table 2
t2.2 Radiocarbon results for Okarito cores

t2.3	Laboratory code	Core	Depth (cm)	dC ¹³	C ¹⁴ age years BP	Calibrated age BP (2 sigma)	Material dated
t2.4	Wk8989	4	65	-26.3±0.2	1800±60	1870–1560	Root from buried soil
t2.5	Wk8987	6	78	-21.5±0.2	2170±70	2340–1950	Bulk soil/organics — buried soil
t2.6	Wk8618	6	82	1.1±0.2	10,580±60	12,280–10,880	Articulated shell — <i>Austrovenus stutchburyi</i>
t2.7	Wk8986	6	82	-0.6±0.2	900±60	630–455	Articulated shell — <i>Austrovenus stutchburyi</i>

t2.8 Calibrations based on [Stuiver et al. \(1998\)](#) using a ΔR of -30 ± 15 for marine shells ([McFadgen and Manning, 1990](#)).

425 level for only one sample (76–78 cm, [Table 1](#)). The fauna
426 from this sample indicate a wide elevation range, from
427 subtidal (2.5 m below MSL) to mean high water neap
428 (1.75 m above MSL).

429 In summary, the forams identified from these cores
430 provide evidence for abrupt changes in salinity during
431 the accumulation of the upper metre of Okarito Lagoon
432 infill. We have confidence in this evidence because the
433 preservation state of tests is moderate to good, there is
434 no evidence for reworking of older tests into the sam-
435 pled sediments and open marine fauna were not re-
436 corded; together indicating an intact and *in situ* faunal
437 assemblage that reliably reflects palaeo-salinity condi-
438 tions at the sample sites.

439 4.5. Radiocarbon ages

440 Radiocarbon results for four samples taken from cores
441 4 and 6 are presented in [Table 2](#). Samples of buried soil
442 and root material from within yielded similar uncalibrat-
443 ed ages of 1800±60 years BP (Wk8989; core 4) and
444 2170±70 years BP (Wk8987, core 6). Two specimens of
445 articulated bivalve, taken from the base of the buried soil
446 in core 6, yielded contrasting ages of 10,580±60 years
447 BP (Wk8618) and 900±60 BP (Wk8986). We view the
448 older age as unreliable on the basis that sea level was
449 about 30 m lower than present at ~10 ka and the area now
450 occupied by Okarito Lagoon would have ice-marginal
451 terrain. Contamination of sample Wk8618 by old carbon
452 is therefore assumed and the age rejected. The true
453 Middle Holocene age for the lower part of the lagoon
454 infill is supported by a C¹⁴ age of 6380±60 years BP on
455 articulated cockle (*A. stutchburyi*) recovered from 326 cm
456 depth in core 6 as part of a previous study ([Goff et al.,](#)
457 [2004](#)).

458 5. Interpretation and discussion

459 The preservation of a buried soil at similar depths
460 below the modern salt marsh and fringing tidal flat of
461 Okarito Lagoon is strong evidence for localized sub-

462 sidence. Results from diatom- and foram-based recon-
463 struction of palaeo-salinity support an interpretation of
464 abrupt changes in local environmental conditions across
465 the contact between the buried soil and overlying de-
466 posits. Thus, we recognize an increase in the salinity of
467 lagoon waters associated with increased water depth of
468 approximately 50 cm. Further evidence for increased
469 tidal incursion at core site 6 is given by geochemical data
470 that record a marked increase in a suite of elements to
471 indicate marine-enriched lagoon waters (especially
472 S, Na, Ca and Ba), associated with sediments directly
473 above the buried soil. We note, however, that these
474 sediments do not contain any open marine foraminifera
475 species, so that evidence for transport of material from
476 the open coast is lacking.

477 Sediments directly overlying the buried soil range in
478 grain size from coarse sand and gravel (trench 1 and 3)
479 to silty fine sand (core 4 and 6, trench 2). Our sampling
480 does not allow for detailed mapping of these deposits,
481 but local variability of grain size such as between core 6
482 and trench 1 (a distance of ~100 m) suggests that
483 sediment texture may vary with proximity to tidal chan-
484 nels, with channels being the likely source for local
485 reworking of coarse-grained sediments. These deposits
486 grade-upward to silty fine sands that contain an increas-
487 ingly dominant freshwater diatom assemblage, which
488 suggests that the site became less open to tidal influence
489 following deposition of the coarser bed. Abrupt declines
490 in geochemical indicators of saline conditions point to a
491 similar transition. On this basis, we interpret the finer
492 sediments as tidal flat deposits, formed under regular,
493 low-energy conditions.

494 Here we propose a scenario to account for the ob-
495 served changes in the shallow stratigraphy of Okarito
496 Lagoon. Given the proximity of Okarito Lagoon to the
497 Alpine fault, the most likely mechanism for subsidence
498 of the lagoon floor is earthquake activity. Co-seismic
499 subsidence has not previously been recognized for the
500 West Coast, but is well established for similar tectonic
501 settings elsewhere; most notably the Pacific Northwest
502 coast of the USA ([Atwater, 1987](#)). Subsidence of coastal

503 lowlands has also been associated with tsunami inun-
504 dation of these coasts, with sedimentary evidence pre-
505 served as anomalous deposits (sand sheets) across the
506 subsided surface (Atwater and Moore, 1992).

507 This leads to the question — did co-seismic subsi-
508 dence of 0.5 m in Okarito Lagoon trigger a local tsunami
509 that left a sedimentary record? Based on the results
510 presented, we highlight the following as potential tsu-
511 nami evidence: (1) the gravel and pebble deposits in
512 trench 1, and to a lesser degree, the coarse sand and fine
513 gravel in trench 3 are a compelling indicator of high-
514 energy flow (metres per second) across the subsided soil
515 surface; (2) the abrupt rise in lagoon salinity as recorded
516 by geochemical data, diatoms and forams is consistent
517 with, though not exclusive to, sudden incursion of
518 marine waters, and; (3) the re-establishment of reduced
519 salinity conditions in sediments above the coarse-
520 grained deposits, as indicated by diatoms particularly,
521 strengthens the interpretation that higher salinities are
522 not the norm for Okarito Lagoon.

523 In evaluating the case for a tsunami in Okarito La-
524agoon (and adjacent coast), we must also consider other
525 scenarios as explanation for the observed sedimentary
526 record. For these alternatives, we draw upon geomor-
527 phic and historic evidence. We noted earlier that the
528 barrier to Okarito Lagoon has a series of blind channels
529 that terminate behind the modern foredune (Fig. 2).
530 These channels are interpreted as former tidal inlets.
531 Given that the current (micro) tidal prism of the lagoon
532 only requires one inlet, we assume that the lagoon has
533 always functioned as a single-inlet system, the location
534 of the inlet has shifted and that at times the lagoon has
535 remained closed. Support for this assumption is found
536 in the raised shoreline bench and shallow caves that
537 fringe the landward side of Okarito Lagoon. The sim-
538 plest explanation for this raised shoreline is that the
539 lagoon was up to 1 m deeper and closed to tidal ex-
540 change. An alternative is that the bench formed during
541 the Middle Holocene sea-level highstand. However, the
542 highstand remains to be convincingly documented for
543 this coast. Moreover, recent dendrochronological work
544 on the rimu (*Dacrydium cupressinum*) and kahikatea
545 (*Dacrycarpus dacrydioides*) forest that has colonized
546 the bench indicates that the oldest trees post-date AD
547 1832 (Goff et al., 2004). Thus, it is likely that low-
548 ering of the lagoon water level occurred shortly prior
549 to ca. AD 1830 and was presumably associated with
550 barrier breaching to re-establish tidal exchange, not a
551 change in relative sea level. Goff et al. (2004) propose
552 tsunami as the mechanism for barrier breaching and
553 identify a known tsunami in AD 1826 as the likely
554 event.

Further indirect evidence for subsidence and tsunami
along this coast is found in historical accounts from the
mid-nineteenth century and is summarized in Goff et al.
(2004). Of these accounts, keys for this study are two
reports of a drowned Maori village at Poherua (aka
Saltwater) Lagoon 11 km to the north of Okarito Lagoon.
The first report was from the explorer Thomas Brunner
in 1848 (republished in Pascoe, 1952) and the second in
1864 by Dobson, who later wrote, “At Poherua lagoon
there had been at one time a Maori village of con-
siderable size, the stumps of the posts of the houses
showing plainly...The land had sunk..., as these traces
could be seen at dead low water, spring tides.” (Dobson,
1930, p.72). Dobson also noted that local Maori were
unaware the village had ever existed, which suggests to
us that its subsidence and abandonment occurred well
before the 19th century. This site was last verified in the
1980’s (New Zealand Forest Service, in press).

Returning to the point regarding lagoon closure, this
will be achieved by accretion of the beach berm on the
barrier, leading to blocking of the inlet (a process that
occurs about once a decade today). Clearly, this requires
a positive sediment budget in the littoral zone and we
recognize the large gravel-bed rivers (Waiho River and
Cook River) to the south of Okarito Lagoon as obvious
long-term sediment sources for barrier construction.
However, it is unlikely that this supply has been at a
consistent rate. There is good evidence from the Haast
coast, ~100 km to the south of Okarito Lagoon, that
barrier construction has been episodic and linked to
fluctuations in sediment discharge from rivers draining
the Southern Alps. Reconstruction of the chronology of
dune-ridge formation by Wells and Goff (2006) has
identified a link to known earthquakes on the Alpine
fault. It follows that these events have led to increased
sediment supply to rivers from earthquake-induced
landslides, and in turn, to the coast. Given this event-
based fluctuation in sediment supply to barriers, perio-
dic closure of lagoons along the West Coast is an ex-
pected outcome.

To summarize, fluctuations in the salinity and water
level of Okarito Lagoon can be readily accounted for by
periodic opening and closing of the tidal inlet entrance.
However, this cannot be used to explain the placement
of gravel and coarse sand at relatively distal locations in
the lagoon, nor the preservation of a buried salt marsh
soil. The shallow depth of the gravel and sand deposits
also precludes transport as a lag along a tidal channel
because flow velocities in a <1 m deep channel inside a
microtidal lagoon would not exceed threshold for gravel
transport. We also discount river floods as a transport
mechanism on the basis that the competence of the

607 Okarito River is limited to transport of coarse sand to the
608 delta front, which is on the opposite shore of the lagoon,
609 ~1.4 km from the site of trench 1. Tsunami remains the
610 only plausible mechanism for transport of coarse sand
611 and gravel to this relatively quiescent location. And the
612 association with a subsided soil provides the important
613 link to palaeo-seismicity.

614 6. Age of this event?

615 Establishing the age for a palaeo-tsunami event is
616 problematic, with the best results often obtained by
617 dating material in sedimentary units above and below an
618 interpreted tsunami deposit, to “age bracket” the event.
619 Radiocarbon ages are often equivocal because of the
620 possible introduction of older reworked carbon (wood
621 and other macro-/microfossil debris) (Dawson, 1994;
622 Goff et al., 1998). We have already discounted one
623 reported age ($10,580 \pm 60$ years BP) on this basis. Ack-
624 nowledging the further possibility that organic material
625 in the buried soil contains old carbon derived from the
626 catchment, the two radiocarbon ages of 1800 ± 60 years
627 BP and 2170 ± 70 years BP are interpreted as maximum
628 ages for deposition of the enclosing sediments. Our
629 remaining radiocarbon age of 900 ± 60 years BP is de-
630 rived from an *in situ* bivalve preserved within the buried
631 soil. The calibrated age of this sample is in the range
632 630–455 years BP (1320–1495 AD) and represents the
633 best age estimate for the subsidence event and tsunami.
634 This age correlates well with the earliest in a series of
635 recognized ruptures on the Alpine fault, dated to ca. AD
636 1450 (Bull, 1996; Yetton et al., 1998). Other ruptures are
637 dated to AD 1620, 1717 and 1826, with the latter gener-
638 ating a tsunami. These events are also correlated with
639 large-scale forest disturbance due to severe ground
640 shaking and landslides (Wells et al., 1998, 1999, 2001;
641 Cullen et al., 2003). It seems highly likely therefore that
642 the tsunami reported here is linked directly to the mid
643 15th century event, either as the direct result of sub-
644 sidence associated with onshore fault rupture, subma-
645 rine fault rupture (the southern segment of the Alpine
646 fault is submarine) or of submarine landsliding gener-
647 ated by an earthquake. These sources remain to be
648 verified, however. It is also suggested that the mid 15th
649 century event caused sufficient forest destruction to
650 trigger increased sediment supply to the coast and, in
651 turn, to closure of the tidal inlet to Okarito Lagoon.

652 7. Conclusion

653 Multiproxy analysis of a distinct sedimentary succes-
654 sion recovered at five sites in Okarito Lagoon provided

for reconstruction of palaeoenvironments that record co- 655
seismic subsidence of a salt marsh surface and inun- 656
dation of that surface by tsunami. Key criteria include a 657
buried soil, associated with major upward changes in 658
grain size, geochemistry, diatom and foraminiferal as- 659
semblages. Limited radiocarbon dating of the buried soil 660
suggests that the subsidence occurred in the mid 15th 661
century and was linked to a rupture on the nearby Alpine 662
fault. Of importance to this case is the necessity to 663
separate other possible mechanisms from the subsi- 664
dence-tsunami interpretation. Thus, geomorphic changes 665
to the barrier inlet are also recognized as an important 666
influence on lagoon water level and salinity, but not as a 667
causal mechanism for the deposits preserved in the la- 668
goon fill. Ultimately, however, there is a link because 669
those changes in barrier form can be potentially traced 670
back to major seismic events in the region. Further work 671
to test this hypothesis will require extension of the record 672
presented here to demonstrate repetition of the process 673
over time and across new study sites. 674

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