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**Article:**

King, Daniel J., Newnham, Rewi M., Gehrels, W. Roland et al. (1 more author) (2020) Late Holocene sea-level changes and vertical land movements in New Zealand. *New Zealand Journal of Geology and Geophysics*. ISSN 1175-8791

<https://doi.org/10.1080/00288306.2020.1761839>

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1 **Late Holocene Sea-Level Changes and Vertical Land Movements in**  
2 **New Zealand**

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# 13 **Late Holocene Sea-Level Changes and Vertical Land Movement in** 14 **New Zealand**

## 15 ABSTRACT

16 Coasts in tectonically active regions face varying threat levels as land subsides or uplifts  
17 relative to rising sea levels. We review the processes influencing relative sea-level change  
18 in New Zealand, and the geological context behind ongoing land movements, focussing  
19 on major population centres. Whilst Holocene sea levels have been reconstructed using a  
20 variety of techniques, recent work uses salt-marsh microfossil assemblages to reconstruct  
21 relative sea-level changes over the past few centuries. For the twentieth century, these  
22 proxy-based studies often show enhanced rates of sea-level rise relative to tide-gauge  
23 observations. The effects of tectonic subsidence must be considered, alongside vertical  
24 and dating uncertainties in the sea-level reconstructions. Global Positioning Systems  
25 (GPS) observations for the past few decades show that vertical land movement (VLM)  
26 may be influencing rates of relative sea-level rise. However, the short period of GPS  
27 observations, during which trends and rates have varied at some localities, raises  
28 questions over the longer-term contribution of VLM to sea-level change over the past few  
29 centuries and for future projections. We argue that high-resolution palaeo-sea-level  
30 reconstructions from salt-marsh sedimentary sequences can help to answer these  
31 questions regarding the interplay between sea-level change and VLM at key locations.

32

33 **KEYWORDS:** Sea level, Holocene, New Zealand, palaeoenvironment, climate change,  
34 sea-level rise, palaeoclimate, vertical land movement, tectonics, palaeoseismicity

## 35 **Introduction**

36 The tectonic complexity of New Zealand introduces a great deal of uncertainty  
37 in the projection of future local sea-level changes. New Zealand's land-use planning  
38 policies typically work on a timeframe of 100 years or so (Department of Conservation  
39 2010; Bell et al. 2017; Ministry for the Environment 2017), with the Ministry for  
40 Environment currently recommending infrastructure guidelines that plan for a minimum

41 1 m of sea-level rise by 2120. During this timeframe, however, the rate of relative sea-  
42 level change will vary considerably from region to region due to New Zealand's  
43 complex tectonics.

44

45 For example, in the capital city Wellington, continuous global positioning  
46 system (cGPS) records reveal that the land is currently undergoing tectonic subsidence  
47 at a rate of 3 mm/yr, although periodic uplift, approximately every five years during  
48 slow slip events (SSEs), reduced this to a net rate of 2.2 mm/yr between 2000 and 2015  
49 (Denys et al. 2017). Assuming the interseismic subsidence and periodic SSE uplift  
50 pattern continues, this region is expected to experience enhanced sea-level rise in the  
51 future relative to regions of tectonic stability, and at present shows considerably faster  
52 rates than other parts of the country (Cole 2010).

53

54 Whilst it is apparent from the Wellington example that vertical land movement  
55 (VLM) needs to be factored into local to regional scale relative sea-level projections in  
56 New Zealand, questions arise from the short period of instrumental observations. The  
57 longest continuous cGPS records only span around twenty years and the longevity of  
58 trends observed in the cGPS record is unknown. In many cases (particularly in the  
59 North Island), the longer term (tens to hundreds of kyr) VLM trends determined from  
60 geological and geomorphological observations are the reverse of the observed cGPS  
61 trend (Beavan and Litchfield 2012; Stephenson et al. 2017). Also, cGPS data are  
62 presented relative to a reference station, which in New Zealand has traditionally been  
63 Auckland, owing to its assumed vertical stability (e.g. Beavan and Litchfield 2012;  
64 Houlié and Stern 2017). Nevertheless, some studies have used the International

65 Reference Frames ITRF2000 (Tenzer and Gladkikh 2014) and ITRF2008 (Tenzer and  
66 Fadil 2016; Denys et al. 2020), allowing for the identification of VLM in any Auckland  
67 cGPS records, and the rest of the country, relative to the Earth's geoid. A limited  
68 number of studies contextualise VLM in a longer-term (Holocene) timescale (e.g.  
69 Hayward et al. 2012; 2015a; 2015b; 2016), but notwithstanding the oft-cited work by  
70 Gibb (1986), knowledge of Holocene relative sea-level changes in New Zealand is  
71 scarce and fragmented (Clement et al. 2016).

72

73         Here we review the present knowledge of Holocene relative sea-level change in  
74 New Zealand to provide geological context and to test the longevity of current  
75 instrumentally observed VLM patterns (Figure 1) with the aim to help better understand  
76 how parts of New Zealand are likely to be affected by future sea-level rise. Common  
77 methods of Holocene sea-level reconstruction are summarised in Figure 2. We focus  
78 especially on the late Holocene (past ~2 kyr) because it provides temporal continuity  
79 with the instrumental era. Also, Gibb (1986) suggested that sea level (i.e. the 'regional  
80 eustatic' signal unrelated to VLM) was approximately stable in New Zealand during the  
81 late Holocene until the last ~100 years. First, we consider the tectonic and climatic  
82 controls on relative sea-level change in New Zealand, followed by evaluation of  
83 previous studies that have generated relative sea-level curves in New Zealand, with  
84 emphasis on those which have used foraminiferal assemblages as sea-level proxies to  
85 create high-resolution centennial-scale records.

86         Owing to its multiple possible meanings, Gregory et al. (2019) decried the  
87 use of the term 'eustatic' or 'global eustasy' to refer to global sea-level changes  
88 and influences. Several of the papers discussed in this review generate 'regional

89 eustatic curves' (defined sensu Gibb (2012) as a time-dependent approximately  
90 uniform change in sea level around New Zealand). This definition is flawed,  
91 because 'eustatic' sea-level change is not uniform around New Zealand due to  
92 oceanographic and gravimetric processes. Here, we use the term 'sea-surface  
93 height' for the changes in water height relative to the Earth's ellipsoid instead. For  
94 all other wordage, this paper primarily uses the sea-level terminology of Gregory  
95 et al. (2019), with relative sea level being defined as the change in local mean sea  
96 level relative to the solid surface (i.e. seafloor or land), where mean sea level is  
97 determined relative to the international reference frame. Spatial variability is  
98 observed in both long-term (Holocene) relative sea-level changes (Clement et al.  
99 2016) as well as in recent sea-level changes during the satellite altimetry era (see  
100 Ackerley et al. 2013) that shows up to 2.9 mm/yr of sea-level rise around the  
101 coasts of Auckland and Northland, decreasing southwards to 2.1 mm/yr around  
102 the coast of the southern South Island since 1993 (AVISO 2019) (Figure 3).

### 103 **Controls on relative sea-level change in New Zealand**

#### 104 *Tectonic controls*

105 Regional tectonics dominate the relative sea-level signature across much  
106 of New Zealand, and trends of uplift and subsidence can vary significantly  
107 depending on the timescale of analysis. For example, while cGPS records show  
108 that much of the east, south, and west coasts of the North Island are aseismically  
109 subsiding at a mean rate of  $\sim 1.5$  mm/yr (Beavan and Litchfield 2012), uplifted  
110 marine and fluvial terraces show that these same regions have undergone  
111 substantial net uplift over the past 125 kyr (e.g. Grapes 1991; Beavan and  
112 Litchfield 2012; Ninis 2018) in response to upper plate and/or subduction

113 earthquakes (Berryman 1993a; 1993b; Berryman et al. 2012; Clark et al. 2015).  
114 Over the longer time scale (the past ~125 kyr), approximately 45% of the New  
115 Zealand coast has been undergoing uplift (with uplift being gradual and aseismic  
116 in northwest North Island, northwest South Island, and Bay of Plenty), 15% has  
117 been undergoing subsidence, associated with the plate boundary zone, and the  
118 remaining 40% has been either stable or lacks sufficient data to determine its  
119 stability (Beavan and Litchfield 2012).

120

121 It is firmly established that the North Island's tectonic deformation is the  
122 result of subduction of the Pacific Plate beneath the Australian Plate at the  
123 Hikurangi Margin over the past 20–25 Myr (Rait et al. 1991). Both the degree of  
124 coupling and the obliquity of convergence between the plates increase southwards  
125 along the North Island (Litchfield et al. 2007; Wallace and Beavan 2010) (Figure  
126 1B), with the regions most strongly coupled undergoing interseismic subsidence  
127 in excess of  $-3$  mm/yr (Houlié and Stern 2017), and in some cases by as much as  
128  $-10$  mm/yr (Tenzer and Gladkikh 2014). Further north (in the central and eastern  
129 North Island), the weakly coupled parts of the margin are undergoing  $1-3$  mm/yr  
130 of uplift (Houlié and Stern 2017). Finite element modelling by Litchfield et al.  
131 (2007) suggests that  $1$  mm/yr of this uplift is the result of the subduction of the  
132 relatively thick and buoyant crust comprising the Cretaceous Hikurangi Plateau  
133 Large Igneous Province. Uplift in excess of  $1$  mm/yr in the northern and central  
134 parts of the margin was explained by their model to be caused by seamount  
135 subduction and sediment underplating. This mechanism explains the lack of

136 faulting linked with uplift of the North Island axial ranges (Houlié and Stern  
137 2017).

138           In the South of the North Island (the Southern Hikurangi Margin), where  
139 coupling-induced interseismic subsidence is gradually lowering the land (Figure  
140 1B) and enhancing relative sea-level rise, cGPS data reveal the presence of slow-  
141 slip events (SSEs): intervals of gradual aseismic slip around the subduction  
142 interface, causing mm-scale uplift in the overlying crust. Since 2003, SSEs have  
143 been documented four times in the south of the Hikurangi margin (Wallace and  
144 Beavan 2010; Wallace et al. 2017; Wallace 2020), with durations between 200  
145 and 480 days and moment magnitudes equivalent to Mw 6.6 (2003) – 7.2 (2004–  
146 5) (Wallace and Beavan 2010). As mentioned in the introduction, cGPS data from  
147 two sites, situated 30 km apart, show that these events have diminished the impact  
148 of subsidence on the expected sea-level trend of Wellington by 0.8 mm/yr  
149 between 2000 and 2015 (Denys et al. 2017, 2020).

150           In the northern Hikurangi Margin, around Gisborne and Hawke’s Bay,  
151 SSEs occur far more rapidly on the shallow portion of the subduction interface  
152 (10–15 km), with a recurrence interval of approximately two years. They are also  
153 of much shorter durations than in the south, typically lasting for approximately  
154 two weeks (Wallace and Beavan 2010). The initiation of these events has been  
155 linked to seamount subduction (Barker et al. 2018; Schwartz et al. 2018). In the  
156 southern Hikurangi margin, SSEs occur deeper (25–60 km) than those in the  
157 northern margin, but both occur in regions of frictional stability, in either the  
158 transition zone between velocity-strengthening and velocity weakening-behaviour  
159 on the plate interface, or in regions of high fluid pressure (Wallace and Beavan



160 2010). The processes causing SSEs are not entirely clear, but it is now known that  
161 they can be triggered by earthquakes from the far-field, with SSEs in both the  
162 northern and southern Hikurangi Margin having been dynamically triggered by  
163 passing waves from the 2016 Mw 7.8 Kaikōura earthquake in the northern South  
164 Island (Wallace et al. 2017, 2018; Bartlow et al. 2018; Wallace 2020). Uplift in  
165 Wellington during this event has accommodated much of the past decade's  
166 subsidence at the Wellington cGPS sites (see GeoNet 2019a). Whether this uplift  
167 should be treated as the amplification of the SSE by post-seismic deformation, or  
168 the uplift is dominated by the postseismic signal (as modelled by Denys et al.  
169 2019), is currently a matter of debate and ongoing research.

170

171         There is currently no geological evidence to evaluate the longer-term  
172 patterns and recurrence times of SSEs in New Zealand, though these will become  
173 clearer with time as cGPS records lengthen. However, in Wellington at least, it  
174 would seem reasonable to assume that large SSEs and similar aseismic uplift  
175 events have occurred multiple times during the last century, if we assume a  
176 constant rate of coupling-induced inter-SSE subsidence (currently 3 mm/yr). If  
177 this were not the case, then relative sea-level rise in Wellington since the 1943  
178 relocation of the Wellington tide gauge would be expected to be  $\sim 3.3$  mm/yr:  
179 comprising  $\sim 1.1$  mm/yr mean rise in sea-surface height around New Zealand for  
180 the twentieth century (Tenzer and Gladkikh 2014) plus 2.2 mm/yr net ground  
181 subsidence (Denys et al. 2017). However, the observed rate over the past century  
182 is  $2.18 \pm 0.17$  mm/yr (Denys et al. 2020), and even less since 1943; considerably  
183 lower than would be expected if the rates of change observed in the pre-2015

184 cGPS record were representative of the entire past century (and less than the  
185 magnitude of relative sea-level rise that should be derived from subsidence alone,  
186 if the cGPS trend were representative of the long-term VLM rate). Whether this  
187 discrepancy between expected and actual rates of relative sea-level rise in  
188 Wellington results from large (and/or frequent) SSEs or fluctuating inter-SSE  
189 subsidence rates is not currently clear. The fact that calculations by Tenzer and  
190 Gladkikh (2014) for New Zealand's changes in sea-surface height questionably  
191 assume no regional variability may also be a source of discrepancy, as this  
192 assumption is not supported by satellite data (Figure 3).

193         One area of note on Figure 1A, where cGPS data show abnormally fast  
194 rates of uplift (9.8 mm/yr relative to ITRF2008 (Tenzer and Fadil 2016)) is the  
195 central Bay of Plenty near Matatā. As discussed by Beavan and Litchfield (2012),  
196 this uplift is predominantly the result of a series of earthquake swarms initiating  
197 around 2005. Prior to this, the heights and ages of nearby raised beaches suggest  
198 that uplift over the past 5 kyr has generally been very slow (0 to ~0.6 mm/yr),  
199 albeit with several instances of abrupt uplift and subsidence (Begg and  
200 Mouslopoulou 2010). Modelling by Hamling et al. (2016) suggests that the  
201 modern earthquake swarms are the result of the inflation of a large magma  
202 chamber at a depth of ~9.5 km, associated with rifting in the Taupo Volcanic  
203 Zone (TVZ). Lamb et al. (2017) gave an alternate explanation for the uplift,  
204 attributing it to melting-induced episodic changes in vertical flow forces  
205 associated with mantle upwelling in the TVZ rift axis. In any case, the abnormally

206 fast rates of uplift near Matatā appear to be a localised and temporary phenomena  
207 related to volcano-tectonic rifting processes.

208 In the South Island, continental collision, transpression around the Alpine  
209 Fault, and crustal thickening cause a general signature of uplift in the cGPS  
210 record, with uplift being greatest proximal to the Alpine Fault (6–8 mm/yr), and  
211 decreasing to 1–2 mm/yr towards the coastal regions (Houlié and Stern 2017).  
212 Most of the coastline to the northeast and southwest of the Southern Alps is  
213 undergoing subsidence, typically on the order of 1.4–1.5 mm/yr in the northwest  
214 South Island (Tenzer and Fadil 2016) and <1 mm/yr in the southeast South Island  
215 (Tenzer and Fadil 2016), although much of this subsidence is within the margin of  
216 error and may not be a genuine signature (Stern 2019 pers. comm.). The central  
217 western South Island may be tectonically stable (Beavan and Litchfield 2012), but  
218 limited data exist to provide any confidence in this. Tenzer and Gladkikh (2014)  
219 note that VLM velocity rates are, on average, notably much faster on Australian  
220 Plate (where stations have a mean VLM velocity of  $-1.4$  mm/yr) than on the  
221 Pacific Plate (mean VLM velocity  $0.5$  mm/yr). However, it should be noted that  
222 the comparative characterisation of vertical plate motion on the Pacific Plate is  
223 possibly less reliable than on the Australian Plate, due to fewer and widely  
224 dispersed Pacific Plate (and South Island) stations (Figure 2), and that VLM is  
225 negligible in several Australian Plate localities (Figure 1A) (furthermore, it may  
226 be difficult to confidently say which plate sites located along the plate boundary  
227 zone around the Alpine Fault belong to). The Dunedin region is of particular note  
228 regarding VLM on the Pacific Plate. Here, subsidence between  $-0.66$  and  $-1.89$   
229 mm/yr is observed (Denys et al. 2020), yet only  $1.35 \pm 0.15$  mm/yr relative sea-  
230 level rise is recorded from the city's tide gauge since 1899 (Denys et al. 2020).

231 This is the lowest rate of relative sea-level rise observed in New Zealand, and  
232 indicates, by extension of the GPS trends across the past century, approximately 0  
233 mm/yr sea-surface height change across some parts of the harbour over the past  
234 century (Denys et al. 2020). Noting that this is unrealistic, Denys et al. (2020)  
235 proposed that the low-rate of long-term sea-level rise is the result of frequent  
236 uplift events associated with earthquakes in Fiordland.

### 237 *Climatic controls*

238 During the Common Era, the main climatic processes contributing to sea-  
239 level change have been ice melt and thermal expansion accompanied by ocean  
240 circulation. Prior to the industrial revolution, global climate in the Common Era  
241 underwent an irregular, long-term cooling trend ( $-1.1$  to  $-0.3^{\circ}\text{C}/\text{kyr}$ ), with  
242 globally asynchronous intervals of heightened cooling or warming (Ahmed et al.  
243 2013; McGregor et al. 2015; Neukom et al. 2019). Global mean sea level  
244 fluctuated by up ca. 0.1 m on multi-decadal to centennial timescales (Kopp et al.  
245 2016). The cooling trend terminated around  $\sim\text{AD } 1800$  with sustained global  
246 temperature rise accompanying increased anthropogenic greenhouse gas  
247 emissions (McGregor et al. 2015). These climate patterns are also apparent in  
248 New Zealand, where archaeological and palynological evidence suggests that the  
249 first Polynesian settlers ( $\sim\text{AD } 1250$ ) encountered a warmer climate than did the  
250 first European settlers ( $\sim\text{AD } 1800$ ) (Anderson 2014; Newnham et al. 2018),

251 although a tree-ring record from Westland suggests that a gradual warming trend  
252 may have initiated around AD 1610 (Cook et al. 2002).

253           From the middle 1800s, melting of ice sheets and glaciers worldwide and  
254 thermal expansion have been the main contributors to sustained global mean sea-  
255 level rise (Church et al. 2013), with an additional contribution from the melting of  
256 glaciers (Leclercq et al. 2014). Over a 40-year period centred on ca. AD 1925,  
257 proxy and measured sea-level datasets from around the world show a significant  
258 positive inflexion, hypothesised to be in association with ice-mass loss in the  
259 Arctic region (Gehrels and Woodworth 2013). Initially only a minor component  
260 of sea-level rise (Wigley and Raper 1987), observations from bathythermographs  
261 and Argo floats now suggest that thermal expansion has increased its contribution  
262 to global sea-level rise from 0.6 mm/yr (1971–2010) to 0.8 mm/yr (1993–2010)  
263 (Church et al. 2013). Tide-gauge records suggest a global mean sea-level rise of  
264 1.1–1.2 mm/yr from 1901–90 (Hay et al. 2015; Dangendorf et al. 2017),  
265 accelerating significantly to  $3.3 \pm 0.4$  mm/yr during the satellite altimetry era  
266 (Cazenave and Remy 2011), and still accelerating at a mean rate of  $0.084 \pm 0.025$   
267 mm/yr<sup>2</sup> (Nerem et al. 2018).

268           These observed sea-level changes for the past century are not globally  
269 uniform, however, owing to a number of climate-related variables, including  
270 oceanic and gravimetric responses to ice melt. Satellite data show that much of the  
271 Tropical and Subtropical Pacific has experienced particularly high rates of rise  
272 since at least 1993 due to the influence of the Interdecadal Pacific Oscillation  
273 (IPO), El-Niño Southern Oscillation (ENSO), and wind forcings (Mimura and  
274 Horikawa 2013). These factors also influence New Zealand (Figure 3), resulting

275 in a range of sea-level projections (all higher than the global average) over the  
276 coming century (Ackerley et al. 2013) (Figure 1A). Between 1993–2019, the  
277 coastlines around parts of Northland and Auckland experienced 4.6–4.8 mm/yr  
278 sea-surface rise, generally declining further south, with 4–4.3 mm/yr sea-surface  
279 rise off the coast of much of the South Island, and 3.8 mm/yr off the coast of  
280 Fiordland. Major exceptions to this trend can be observed around the Greater  
281 Wellington region, with 4.8–5.2 mm/yr, and around the northern South Taranaki  
282 Bight and the Tasman Bay, where sea-surface rise as low as 1–2 mm/yr is  
283 measured (Figure 3). The general southwards decrease in this absolute  
284 (geocentric) sea-level rise can possibly be attributed to the equatorward increase  
285 in temperature and thermal expansion, as well as the influence of Antarctica  
286 (where melting ice decreases the gravitational pull of the ice sheet, leading to  
287 redistribution of water mass in the far-field (i.e. closer to the equator), rather than  
288 at proximal high latitudes, as discussed in Tamisiea et al. (2003)).

289         The IPO is a 20–30-year cycle defined by positive and negative phases of  
290 sea-surface temperature (SST) anomalies. Positive phases involve cooler than  
291 normal SSTs in the West Pacific and warmer SSTs in part of the East Pacific.  
292 Negative phases involve warming in the West Pacific and cooling in part of the  
293 East Pacific. Because of this West Pacific warming, negative IPO phases  
294 associated with higher sea surfaces in New Zealand can be identified in New  
295 Zealand's tide-gauge record from 1947–1975, and 1998-present, with an

296 intervening positive phase 1976–1997 (Bell and Hannah 2012). Across a cycle,  
297 the IPO can affect New Zealand’s sea surface by  $\pm 5$  cm (Dawe 2008).

298           Superimposed on the IPO, the ENSO is a recurring climatic pattern in the  
299 Central and East Pacific that occurs over an irregular 2–5 year cycle. The ENSO  
300 is characterised by ‘El Niño’ phases and ‘La Niña’ phases when SSTs, in the  
301 central and eastern tropical Pacific are anomalously warmer and cooler,  
302 respectively. This cycle can affect New Zealand sea surfaces by  $\pm 6$  mm, with rises  
303 during La Niña and falls during El Niño (Hannah and Bell 2012). New Zealand  
304 tree-ring records indicate that the strength of the ENSO increased significantly  
305 during the twentieth century relative to the preceding five centuries, suggesting  
306 that ENSO activity, at least proximal to New Zealand, increases with global  
307 warmth (Fowler et al. 2012).

308           Longer-term Common Era periods that may have affected sea level in the  
309 Pacific include the Medieval Warm Period (MWP) and Little Ice Age (LIA). The  
310 spatial variation in the timing and amplitude of temperature change in these events  
311 leads to a great deal of uncertainty regarding whether or not they are even  
312 expressed in the Pacific, and how they could have impacted regional sea level,  
313 with significant debate around the global or regional nature of the MWP (Hughes  
314 and Diaz 1994; Broecker 2001; Nunn 2007a; Ahmed et al. 2013; Chen et al. 2018)  
315 being largely unresolved. Neukom et al. (2019) noted that several climate  
316 reconstructions across the Common Era do not fit the standard MWP/LIA  
317 narratives, and using data from 257 palaeoclimate proxies plotted on a global grid,  
318 found that <50% of the globe shows consistent timings of cold or warm intervals  
319 in the pre-industrial Common Era. Neukom et al. (2019) showed that regionally

320 specific mechanisms controlled multi-decadal climatic variability prior to the  
321 industrial era, and that neither the MWP nor LIA can be treated as globally  
322 consistent events.

323         Such regional variability in late Holocene climate is apparent in the Pacific  
324 region. Nunn (2007a) attributed Pacific archaeological and palynological evidence  
325 for warming ~AD 750–1250 and cooling ~AD 1350–1800 to the MWP and LIA,  
326 respectively. These timings have been contested, however, with coral  
327 palaeothermometry indicating a relatively warm West Pacific during Nunn’s  
328 postulated LIA (Cobb 2002), and some glacial advances have been identified  
329 during the postulated MWP (Schaefer et al. 2009). Nevertheless, Nunn (2007a)  
330 speculatively linked the Pacific MWP to sea levels ~1 m above the modern in  
331 several sites, such as Lord Howe Island in the Tasman Sea, as well as at Bering  
332 Island in the Bering Sea, and Kunashir Island (North of Japan). These +1 m  
333 horizons are far from globally ubiquitous, however, and there is an overall  
334 scarcity of global high-resolution sea-level data for this time period (Kemp et al.  
335 2011). Indeed, several studies have called Nunn’s conclusions on this matter into  
336 question (e.g. Gehrels 2001; Allen 2006; Fitzpatrick 2010; Clark and Reepmeyer  
337 2012). A cooling interval is present in New Zealand tree-ring records from AD  
338 ~1240–1310 (Cook et al. 2002), coinciding with Nunn’s (2007a) AD 1300  
339 cooling event. However, whether or not such a cooling event was Pacific-wide, or  
340 if the 0.7–0.8 m sea-surface fall postulated by Nunn (2007b) (from evidence from  
341 Pacific Islands) occurred in New Zealand, is far from clear.

342         Longer-term climate-related processes that can affect relative sea-level  
343 change typically relate to the displacement of the lithospheric mantle due to the



344 loading and unloading of ice onto the land during major glacial-interglacial  
345 cycles, in what is referred to as glacial-isostatic adjustment (GIA) (e.g. Benn and  
346 Evans 2010; Whitehouse 2018). Although this is typically regarded as a localised  
347 process due to greater degrees of VLM occurring in proximity to major ice sheets  
348 and glaciers as they melt (or build), the Earth's viscoelastic response to ice-mass  
349 changes is now understood to be manifested globally (Riva et al. 2017). As a  
350 result of far-field continental ice-sheet displacements alone, much of the Northern  
351 Hemisphere is generally undergoing uplift, while much of the Southern  
352 Hemisphere is undergoing subsidence. An area of crust spanning from the  
353 southern Indian Ocean, across the entirety of Australia and New Zealand, and  
354 extending eastwards into the central south Pacific has undergone  $-0.6$  to  $-0.8$   
355 mm/yr of vertical deformation during 2003–14 (Riva et al. 2017). Riva et al.  
356 (2017) note that the rates of far-field vertical deformation vary globally (and,  
357 usually, accelerate) with the increasing acceleration of ice melt, even during the  
358 twentieth and twenty-first centuries.

359         As the effects of longer-term GIA have not yet been quantified for New  
360 Zealand, most palaeo-sea-level and VLM studies apply the global model of Peltier  
361 (2004). This model suggests that much of New Zealand should be undergoing  
362  $\pm 0.1$  mm/yr VLM due to GIA, but fails to accurately account for any of New  
363 Zealand's local ice-mass changes (Cole 2010; Fadil et al. 2013). It may be  
364 beneficial for future studies to use the more recent ICE-6G GIA model (Argus et  
365 al. 2014; Peltier et al. 2015, further improved by Peltier et al. 2018). Work to  
366 quantify the effects of GIA more accurately in New Zealand may prove invaluable  
367 to future sea-level studies. For example, Riva et al's. (2017) model of the  
368 influence of twentieth to twenty-first century ice wastage on VLM was applied to

369 New Zealand's tide gauges by Denys et al. (2020), who showed that it had driven  
370 ~30 mm (or 0.25 mm/yr) subsidence around the South Island, and ~36 mm (or  
371 0.30 mm/yr) subsidence in the North Island (with uplift in the Southern Alps  
372 accounting for the reduced rate in the South Island).

373 GIA also affects global mean sea level indirectly, as the collapse of glacial  
374 forebulges in high-latitude once-glaciated regions prompts the migration of water  
375 into those regions (referred to as 'ocean syphoning') (Mitrovica and Milne 2002).  
376 The loading of additional seawater onto continental shelves can also trigger  
377 upwarping of the adjacent land ('continental levering'; Mitrovica and Milne  
378 (2002)). The process of crustal loading due to the distribution of ocean water is  
379 referred to as hydro-isostasy (Benn and Evans 2010) and its effects on New  
380 Zealand sea level are also currently under investigation, though this process  
381 appears to have been driving subsidence in the Northland Peninsula, on the order  
382 of 1–12 metres across the Holocene, with the magnitude of subsidence increasing  
383 northwards (Clement et al. 2016).

#### 384 ***Orbital Controls***

385 Another mechanism that affects the sea surface in the Pacific on a  
386 centennial time-scale is Earth's rotation. A relationship between the Earth's  
387 rotation and sea level has long been understood, primarily with regard to the  
388 influence of glacioisostatic rebound and water mass distribution on the Earth's  
389 oblateness, and the resultant effect on the rotational state of the planet's orbit (e.g.  
390 Peltier 1988; Nakada and Okuno 2003; Peltier and Luthcke 2009; Mitrovica et al.  
391 2015). This relationship is not in one direction however, as the position of the  
392 Earth's rotational axis affects the position of its 'rotational bulge', which also

393 affects the Earth's oblateness, and from there the distribution of water on its  
394 surface (Mitrovica et al. 2005). It is now also understood that, on a local scale, the  
395 direction by which water from major rivers flows into the ocean is controlled by  
396 the Earth's rotation, leading to elevated water levels to the left of Southern  
397 Hemisphere rivers, though the long-term centennial effects of such processes are  
398 not yet clear (Piecuch et al. 2018)).

### 399 **Holocene sea-level reconstructions**

400 Most studies of Holocene relative sea-level change in New Zealand have  
401 focussed on regions of presumed tectonic stability in order to understand the regional  
402 signature of sea-level rise and use these as a benchmark for work involved in  
403 calculating palaeo-VLM. Estimates of long-term site stability have been based on the  
404 position of the last interglacial shoreline, which was approximately 5 m above modern  
405 sea surface height in New Zealand (Pillans 1990). Gibb (1986) generated a widely-cited  
406 regional 'eustatic' curve for New Zealand based on 82 radiocarbon-dated sea-level  
407 indicators (mostly intertidal molluscs) from presumed tectonically stable sites at  
408 Blueskin Bay, Weiti River Estuary and Kumenga, as well as sites of known instability  
409 such as Pauatahanui, Christchurch, and Firth of Thames. Gibb (1986) isolated tectonic  
410 movements from the presumed unstable localities using data from the presumed stable  
411 sites to remove anomalous data, and generated the curve displayed in Figure 4A. Gibb's  
412 1986 New Zealand record shows a relative sea-level rise from  $-33.5 \pm 2$  m at 10 ka BP,  
413 to approximately the present mean sea level at  $6.5 \pm 0.1$  ka BP. This rise was interrupted  
414 by stillstands during 9.2–8.4 ka BP and 7.5–7.3 ka BP (at  $-24 \pm 2.9$  and  $-9 \pm 2.8$  m  
415 relative to modern sea level, respectively). The 6.5 ka plateauing of sea level reported  
416 by Gibb is approximately coincident with the 7 ka final deglaciation of the Laurentide  
417 Ice Sheet, and a change in global sea-level behaviour from a dominant glacioeustatic

418 control to a dominant glacioisostatic control (Dlabola et al. 2015). Gibb (1986) also  
419 noted decimetre-scale fluctuations in sea level during the past 6.5 ka, identifying a  
420 regression minimum of  $-0.4$  m at 4.5 ka BP, and a transgression maximum of 0.5 m at  
421 3.5 ka BP.

422

423         The Gibb (1986) record has recently come under reconsideration owing to the  
424 low precision of the palaeo-depth indicators and the fact that the radiocarbon dating was  
425 not calibrated conventionally, so cannot be converted into true sidereal years (Clement  
426 et al. 2016). Kennedy (2008) argued that the absence of a mid-Holocene highstand  
427 (observed in other Southwest Pacific relative sea-level reconstructions such as Nunn  
428 (1990) and Baker et al. (2001)) in the Gibb (1986) record may have been due to the  
429 low-resolution of the data, while Clement et al. (2016) suspected the highstand was  
430 assumed to represent uplift and removed from the curve. Doubts were further cast on  
431 aspects of the Gibb curve by Dlabola et al. (2015), who generated a relative sea-level  
432 curve for Fiordland over the last 18 kyr. This Fiordland record, reconstructed using the  
433 heights of overtopped isolation basin sills, as well as diatom assemblages from sediment  
434 cores taken from two fjords, contains a significant increase in the rate of sea-level rise  
435 beginning  $\sim 9.7$  ka BP at a time when Gibb (1986) postulated a stillstand. This record  
436 also indicated a slower initial rate of sea-level rise from 11.4–9.7 ka BP than the Gibb  
437 record. However, Dlabola et al. (2015) acknowledge that the differences may be due to  
438 VLM. What is clear from this discussion is that key assumptions regarding tectonic  
439 stability that underpinned Gibb's seminal New Zealand sea-level record need to be  
440 revisited in light of subsequent studies of sea-level changes and vertical land  
441 movements.

442

443 Hayward et al. (2010, 2015a, 2016) used preliminary New Zealand Holocene  
444 relative sea-level data from an unpublished conference abstract, supplemented by data  
445 from Hicks and Nichol (2007), Dougherty and Dickson (2012), Hayward (2012), and  
446 Hayward et al. (2007, 2012), to infer VLM in several coastal sites across New Zealand.  
447 As with Gibb's (1986) record, this sea-level reconstruction assumed uniform sea-  
448 surface height across all stable coasts of New Zealand, and that all changes thereof were  
449 similarly uniform in magnitude during the middle and late Holocene. This assumption  
450 of regional coherence in timing and amplitude of sea-surface changes is problematic, for  
451 reasons already discussed. Sea-surface change over hundreds of kilometres is seldom  
452 uniform (see Lewis et al. 2013; Clement et al. 2016), as is now evident from the modern  
453 satellite data (Figure 3).

454

455 Clement et al. (2016) addressed the problem of regional variations in relative  
456 sea-level history by integrating a broad selection of mostly published pre-existing local  
457 sea-level proxy data to generate a series of relative sea-level curves for different parts of  
458 New Zealand. They also deployed more advanced radiocarbon techniques than were  
459 available to Gibb (1986), applied GIA corrections using the ICE-5G ice model and  
460 VM2 radial viscosity profile of Peltier (2004), and incorporated feedbacks between GIA  
461 and Earth's rotation (Mitrovica et al. 2005) with consideration for time-dependent  
462 migration of the shoreline (Mitrovica and Milne 2003). In their study, a highstand in the  
463 northernmost North Island was identified from 8.1–7.3 ka BP (0.6–1.4 kyr prior to Gibb  
464 (1986) in agreement with Australian records (e.g. Horton et al. 2007; Lewis et al.  
465 2013)), reaching ~2.65 m above present mean sea level, before falling to present values  
466 between 7.8 and 6.4 ka. Importantly, this highstand was not temporally uniform across  
467 New Zealand, but occurred later further south (as did the first occurrence of present

468 meal sea level). For example, in the South Island, the relative sea level high stand  
469 reached ~2 m above modern sea level from 7.0–6.4 ka BP (Figure 4B-D).

470

471           These regional variations have yet to be fully explained, but Clement et al.  
472 (2016) offered several suggestions, including a decrease in the gravitational attraction of  
473 the shrinking Antarctic Ice sheet, possibly combined with post-glacial meltwater  
474 loading and hydro-isostatic levering. However, as the authors acknowledge, this  
475 hypothesis is not supported by Australian Holocene relative sea-level reconstructions  
476 (e.g. Horton et al. 2007; Lewis et al. 2013), nor the GIA model predictions.  
477 Furthermore, the manner in which GIA was quantified and accounted for by Clement et  
478 al. (2016) calls aspects of the study into question, as the ice model (Peltier 2004) used in  
479 this study has been argued by Cole (2010) and Fadil et al. (2013) to be unsuitable for  
480 use in New Zealand, due to its failure to incorporate isostatic shifts arising from New  
481 Zealand ice-mass changes (as discussed earlier). It may also be questioned how well the  
482 model encapsulates the lithospheric heterogeneities present at the Pacific-Australian  
483 plate boundary, as such issues are not addressed by Peltier (2004) in his discussion of  
484 the model. This particular concern is underpinned by the extreme sensitivity of South  
485 Island glaciers to climatic shifts (Vargo et al. 2017) and their well-documented patterns  
486 of significant growth and deglaciation during the Holocene (e.g. Schaefer et al. 2009).  
487 The possibility, posed by Mathews (1967), that Holocene glacial mass changes in New  
488 Zealand have been sufficient to produce significant GIA requires further investigation,  
489 as does the presumed millennial-scale and short-term vertical stability of much of New  
490 Zealand's coastline, variations in which may contribute to the regional differences  
491 observed in Clement et al.'s (2016) reconstruction.

492

493 **Late Holocene Centennial-Scale Records**

494 Several studies have used benthic foraminiferal assemblages from salt-marsh  
495 sediments to reconstruct Holocene relative sea-level changes in New Zealand in  
496 presumed tectonically stable settings (e.g. Figueira 2012; Gehrels et al. 2008; Grenfell  
497 et al. 2012), or to constrain past vertical land movement in unstable regions (e.g.  
498 Hayward et al. 2004, 2007, 2015a, 2015b, 2016; Clark et al. 2015). Salt-marsh  
499 foraminifera are highly useful in sea-level reconstructions as they occupy far narrower  
500 vertical ranges in the intertidal zone than the bivalves used in previous low-resolution  
501 studies (e.g. Gibb 1986; 2012). This allows sea level to be estimated to within  $\pm 5$  cm  
502 precision in some cases (Southall et al. 2006).

503

504 A key study of this type was conducted at Pounaweia, southern South Island  
505 (Figure 1; Gehrels et al. 2008). The site was considered to be tectonically stable based  
506 on the +4 m height of the Last Interglacial shoreline, as well as local stratigraphic work  
507 by Hayward et al. (2007), which indicated no VLM over the past 1 kyr. The Pounaweia  
508 sea-level reconstruction shows a gradual ( $0.3 \pm 0.3$  mm/yr) relative sea-level rise from  
509 AD 1500–1900, followed by a dramatic increase to  $2.8 \pm 0.5$  mm/yr in the twentieth  
510 century (Figure 5), much higher than the observed twentieth-century New Zealand  
511 mean, estimated as between 1.46 mm/yr (Fadil et al. 2013) and 1.6 mm/yr (Hannah  
512 2004). The high rate of relative sea-level rise at Pounaweia was attributed to a possible  
513 regional high in thermal expansion (Gehrels et al. 2008) although this is not reflected in  
514 tide gauge records for the region. As shown in Figure 5, the sea-level rise at Pounaweia  
515 is notably faster than the sea-level change observed from the nearest tide-gauge records  
516 at Lyttelton, Bluff, and Dunedin (with trends of 2.0, 1.8, and 1.3 mm/yr, respectively  
517 (Hannah and Bell 2012)). However as shown in Figure 5, the spread of the data at

518 Dunedin is far greater than in the other South Island tide gauges, suggesting another  
519 regional influence may apply at that location. Denys et al. (2020) argue that the overall  
520 lower observed rate of sea level rise at Dunedin may reflect the influence of uplift from  
521 earthquake events in Fiordland.

522         Fadil et al. (2013) argued that the high rate of relative sea-level rise  
523 recorded at Pounaweia was due to sediment compaction, which they argued was  
524 poorly constrained in New Zealand. However, Brain et al. (2012) applied  
525 geotechnical modelling experiments to samples from salt marshes in a variety of  
526 depositional settings, and showed that well-consolidated, sub-0.5 m thick salt-  
527 marsh sediment sequences such as those in New Zealand are likely to have only  
528 negligible compaction (on the order of millimetres). It seems therefore that  
529 relative sea-level rise in the southern South Island is noticeably faster than the  
530 regional change in sea-surface height for the majority of New Zealand (even as  
531 observed in the modern satellite record, Figure 3, which shows a post-1993 sea-  
532 level rise of 4.0 mm/yr around Pounaweia), and typical of the rest of the South  
533 Island. Several possible explanations for these higher rates at Pounaweia emerge  
534 from the range of tectonic and climatic controls on New Zealand sea level  
535 discussed earlier in this review. Of these, a possible explanation is localised  
536 aseismic tectonic subsidence in locations where this has not previously been  
537 observed in the geological record. Another possibility is that the relatively small  
538 modern training set developed at the site has constrained the accuracy of the sea-  
539 level reconstruction.

540         At Puhinui Inlet, Auckland (which was assumed to be tectonically stable,  
541 Figure 1), a similar study was conducted by Grenfell et al. (2012), with two sea-



542 level reconstructions (from cores ‘Puh3’ and ‘Puh5’) giving estimates of  $2.8 \pm 0.5$   
543 and  $3.3 \pm 0.7$  mm/yr since 1890. Although these rates are consistent with those at  
544 Pounaweia (Gehrels et al. 2008), they are inconsistent with the Auckland tide-  
545 gauge record, which gives 1.41 mm/yr of rise for this same interval (Cole 2010).  
546 This is problematic, as Denys et al. (2017) find no evidence of a tectonic influence  
547 on the sea-level record for Auckland. One possible explanation relates to the fact  
548 that Puhinui Inlet lies on the west (Tasman Sea) coast of Auckland, whilst the  
549 Auckland tide gauge is situated on the east (Pacific) coast. The difference could  
550 be explained by enhanced thermal expansion on the Tasman Sea side associated  
551 with the region’s ocean circulation pattern. However, if that were the case, it is  
552 not evident in the satellite altimetry record, which shows a 4.6–4.8 mm/yr rise in  
553 sea-surface height adjacent to both Auckland’s Pacific and Tasman coastlines  
554 since 1993 (Figure 3).

555         Alternatively, it is possible that the Puhinui cores were taken from too low  
556 in the salt marsh to enable accurate sea-level reconstructions. This is supported by  
557 the low (<20%) abundance of the high-marsh (Southall et al. 2006) foraminifer  
558 *Trochammina salsa* in the upper part of the cores. As discussed by Hayward et  
559 al. (2016), and Scott and Medioli (1978), palaeoelevation estimate precision  
560 increases significantly (by magnitudes of tens of cm) higher up the marsh, due to  
561 the smaller elevation ranges of foraminifera in the high-marsh environment, as  
562 well as the tendency for sedimentation rates to more closely reflect sea level  
563 (Gehrels and Kemp In press). Uncertainties with the Grenfell et al. (2012) transfer  
564 function may also contribute to the discrepancy, as some of the residual errors are  
565 poorly modelled (+12 and –20 cm), and a negative trend is observable in the  
566 residual errors with respect to increasing height, suggesting a possible

567 overprediction in height for lower elevations, and an underprediction for higher  
568 elevations. As at Pounaweia, the predictive power of the transfer function is further  
569 limited by the small number of surface samples that were taken from Puhinui Inlet  
570 and a significant gap between 275 and 325 cm in the vertical range. These are  
571 common limitations and elsewhere it has been shown that the predictive power of  
572 sea-level reconstructions can be enhanced by developing regional microfossil  
573 training sets that incorporate data from other salt marshes (e.g. Horton and  
574 Edwards 2005; Watcham et al. 2013; Hocking et al. 2017).

575           However, VLM provides another possible explanation for the high rate of  
576 sea-level rise reconstructed for Auckland, despite previous work and assumptions  
577 to the contrary. A 2003–2007 Envisat time series across Auckland reveals three  
578 distinct regions of subsidence within the city (all ~4 mm/yr), with scattered  
579 regions of uplift (maximum of 4 mm/yr) (Samsonov et al. 2010). This land  
580 movement was linked to fluctuations in groundwater recharge and depletion, due  
581 to a lack of correlation with known faults or volcanic centres (Samsonov et al.  
582 2010), an interpretation supported by subsequent genetic algorithm inversion  
583 modelling (Latimer et al. 2010), though not entirely confirmed. Altamimi et al.  
584 (2016) incorporated Auckland’s cGPS sites into their international terrestrial  
585 reference frame (ITRF2014) for global geodetic data, and found the city to be  
586 subsiding by 0.8 mm/yr. Subsidence in Auckland of 0.83 mm/yr relative to the  
587 previous ITRF (ITRF2008) was calculated by Tenzer and Fadil (2016), and  
588 ongoing work suggests that the true subsidence rate is as high as 1.2 mm/yr  
589 (Hreinsdóttir 2019, pers. comm). Whether this subsidence is a long-term process  
590 (it may relate to a viscoelastic response to ice wastage (Riva et al. 2017) and/or to  
591 tectonics), it is of concern because studies of New Zealand cGPS signatures such

592 as Beavan and Litchfield (2012) and Houlié and Stern (2017) assume complete  
593 stability in Auckland, and give all cGPS data relative to this site, meaning that  
594 sites of equivalent subsidence might not be acknowledged in the current literature.  
595 Furthermore, the observed spatial variability in this land movement (Latimer et al.  
596 2010; Samsonov et al. 2010) could provide an explanation for why the proxy sea-  
597 level record at Puhinui differs so significantly from the Auckland tide-gauge  
598 record. Further salt-marsh records may be needed in the Auckland region,  
599 including from higher in the Puhinui salt-marsh sequence, to assess whether or not  
600 the Grenfell et al. (2012) record truly reflects an enhanced relative sea-level rise  
601 due to land subsidence. If the enhanced subsidence is a genuine signal, it would be  
602 ideal to extend the record beyond 1890, in order to assess whether this VLM is  
603 due to anthropogenic disturbance of groundwater (as interpreted by Samsonov et  
604 al. 2010) , or has a longer-term, tectonic origin. Tenuous evidence for long-term  
605 subsidence can be interpreted from Thorne Bay, northern Auckland, where a  
606 wave-cut platform linked with the Last Interglacial is located 3 m above sealevel  
607 (Ballance and Williams 1992), a time when sea level in the region was estimated  
608 to be ~7 m above present (Beavan and Litchfield 2012), although such subsidence  
609 in this locality would be orders of magnitude less than is indicated by Samsonov  
610 et al. (2010)'s data. This difference could result from the extreme spatial  
611 variability in the latter study's observed VLM changes.

612           Figueira (2012) generated sea-level reconstructions from salt marshes at  
613 Waikawa Harbour, southeast Southland, and Whanganui Inlet, northwest Nelson  
614 (Figure 1). These records are by far the longest salt marsh sea-level  
615 reconstructions to date from New Zealand, extending across most of the past  
616 millennium. The record from Waikawa Harbour displays sea-level rise of

617 approximately 2.6 mm/yr between AD ~700 and ~1150, reaching approximately  
618 80 cm above present mean sea-level before abruptly declining to lower-than-  
619 present values by AD ~1250, reaching -40 cm by 1700. A subsequent abrupt  
620 rising trend commencing around 1900 is coincident with the onset of  
621 anthropogenic enhancement. The magnitude of Waikawa reconstructed sea-level  
622 rise between AD ~700 and ~1150 is comparable with speculative estimates from  
623 other Pacific sites for the MWP (Nunn 2007a), but the rate of rise far exceeds  
624 what would be expected from this locality (assuming tectonic stability), while the  
625 termination of the rise and abrupt fall (the 'AD 1300 Event') occurs at least a  
626 century earlier than elsewhere (Nunn 2007a). This high sea-level interval has very  
627 wide error bars, in part because the assemblages used to make this calculation are  
628 rich in the benthic foraminifer *Miliammina fusca*, which has a large vertical range  
629 at the site, and the peak appears to be highly model-dependant. Indeed, the high  
630 abundance of *M. fusca* during this older interval could indicate that the marsh had  
631 colonised a pre-existing mudflat (see Hayward et al. 1999). Immature, flat-  
632 colonising marshes typically have sedimentation rates in excess of the rate of sea-  
633 level rise, creating regressive sequences that are difficult to interpret in terms of  
634 sea-level changes (Gehrels and Kemp In press).

635         Therefore, it is entirely possible that the 'MWP' sea-level peak presented  
636 by Figueira (2012) from Waikawa Harbour is simply an artefact of both the  
637 uncertainty introduced by the assemblage and the problems encountered when  
638 trying to reconstruct sea level using flat-colonising marsh sediments. A mean rate  
639 of sea-level rise of  $3.5 \pm 0.5$  mm/yr is indicated in the Waikawa Harbour record  
640 for the past 120 years, notably greater than any of the South Island tide gauges.  
641 Statistically, far greater confidence is placed on the reconstruction in this part of

642 the record due to its foraminiferal makeup. The Figueira (2012) record from  
643 Whanganui Inlet shows a mean rate of modern sea-level rise of 0.6 mm/yr from  
644 1840–1910, and  $3.6 \pm 0.6$  mm/yr post-1910, also notably faster than the South  
645 Island tide gauges.

646 In summary, detailed sea-level reconstructions for the past few centuries  
647 from different regions (Auckland, northern South Island, eastern Otago,  
648 Southland) all show rates of sea-level rise that are higher than expected from the  
649 nearest tide-gauge records. Clearly, errors and uncertainties in the reconstruction  
650 methodology cannot be ruled out at the sites, but other factors should be  
651 considered as well. If we accept from the work by Brain et al. (2012) that these  
652 sites should experience negligible sediment compaction, it is possible that slow  
653 subsidence has enhanced modern anthropogenic rates of sea-level rise at some of  
654 these sites. In the case of Figueira's (2012) sites, while slow subsidence of  
655 approximately identical rate at both the northern and southern South Island may  
656 seem too coincidental to be realistic, subsidence is entirely consistent with the  
657 vertical cGPS records from Golden Bay and Mahakipawa Hill, in the  
658 Northernmost South Island near Whanganui Inlet (GeoNet 2019b, 2019c), while  
659 subsidence of  $\sim 1.5$  mm/yr in the northwest South Island, proximal to Whanganui  
660 Inlet, was observed by Tenzer and Fadil (2016), who integrated the local cGPS  
661 data into the ITRF2008 global reference frame. Intervals of subsidence and uplift  
662 of approximately equal magnitude are also documented in the cGPS record from  
663 Bluff (the closest cGPS record to both Waikawa Harbour and the Pounaweia site  
664 of Gehrels et al. 2008) (GeoNet 2019d). However, we point out that the vertical  
665 responses to Fiordland earthquakes may differ between Bluff and the Waikawa  
666 and Pounaweia salt marshes, due both to increasing distance from the source, and

667 because the salt-marsh sites lie atop different tectonostratigraphic terranes with  
668 different basement geology to Bluff Harbour (detailed in King 2000).

## 669 **Conclusions**

670 Vertical land movement (VLM) needs to be considered as a factor in all  
671 New Zealand relative sea-level reconstructions, particularly at the multi-  
672 centennial scale where slow tectonic motion may be less obvious than in longer  
673 records. Even at presumed stable locations, VLM could explain enhanced or  
674 dampened rates of sea-level rise observed at the centennial scale. VLM could also  
675 help explain some of the contentious or variable observations from Holocene sea-  
676 level reconstructions in New Zealand and throughout the Pacific. The  
677 consequences of VLM-enhanced relative sea-level rise could be significant for  
678 large cities, most of which are situated on the coast. Of particular concern are  
679 Wellington, where the long-term effects of coupling-induced subsidence and SSEs  
680 are unknown, and Auckland, where satellite altimetry (Latimer et al. 2010;  
681 Samsonov et al. 2010) and high rates of sea-level rise in the Puhinui salt-marsh  
682 record (Grenfell et al. 2012) invite new questions regarding the city's tectonic  
683 stability. Dunedin, Napier and Christchurch are other cities where vulnerability to  
684 VLM-enhanced sea-level rise should be considered. Of particular note, Dunedin's  
685 cGPS record indicates variable subsidence rates up to  $1.89 \pm 0.56$  mm/yr (Denys  
686 et al. 2020), whilst in South Dunedin, around 11,500 people live on lowlands  
687 reclaimed from coastal marshes and dunes, that could prove extremely vulnerable  
688 to future sea-level rise (Morris 2008). Centennial salt marsh sea-level records  
689 from these regions are currently either lacking or have wide uncertainties, and it is  
690 recommended that they are implemented in future work to contextualise trends in

691 modern VLM, as they bridge a gap in scale between high-precision short-term  
692 instrumental records and long-term geological reconstructions.

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1066 Acknowledgements, avoiding identifying any of the authors prior to peer review

1067 This review is a contribution to the NZ SeaRise Programme, funded by the Ministry of Business,  
1068 Innovation and Employment. Special thanks are given to Dr. Alastair Clement and Dr. Paul  
1069 Denys for their thorough and extremely helpful reviews of the manuscript. We also thank Dr.  
1070 Ed Garrett (University of York), for his comments on the manuscript, Dr. Marta Marcos  
1071 (University of the Balearic Islands) for help with generating Figure 3, and Dr. Andrew Rees  
1072 (Victoria University of Wellington) for assistance with transfer functions. Further thanks are  
1073 given to Dr. Sigrún Hreinsdóttir (GNS Science) for provision and discussion of unpublished cGPS  
1074 time series , Prof. Edward Cook (Columbia University) for providing the primary data from the  
1075 2002 tree-ring climate reconstructions, as well as Dr. Paolo Stocci (NIOZ), Dr. Jeremy Gibb  
1076 (Coastal Management Consultancy Ltd., retired), Dr. Susan Schwartz (UC Santa Cruz), Dr. Laura  
1077 Wallace (GNS Science), Prof. Simon Lamb (Victoria University of Wellington), and Prof. Tim  
1078 Stern (Victoria University of Wellington), for valuable conversations about aspects of the  
1079 literature.

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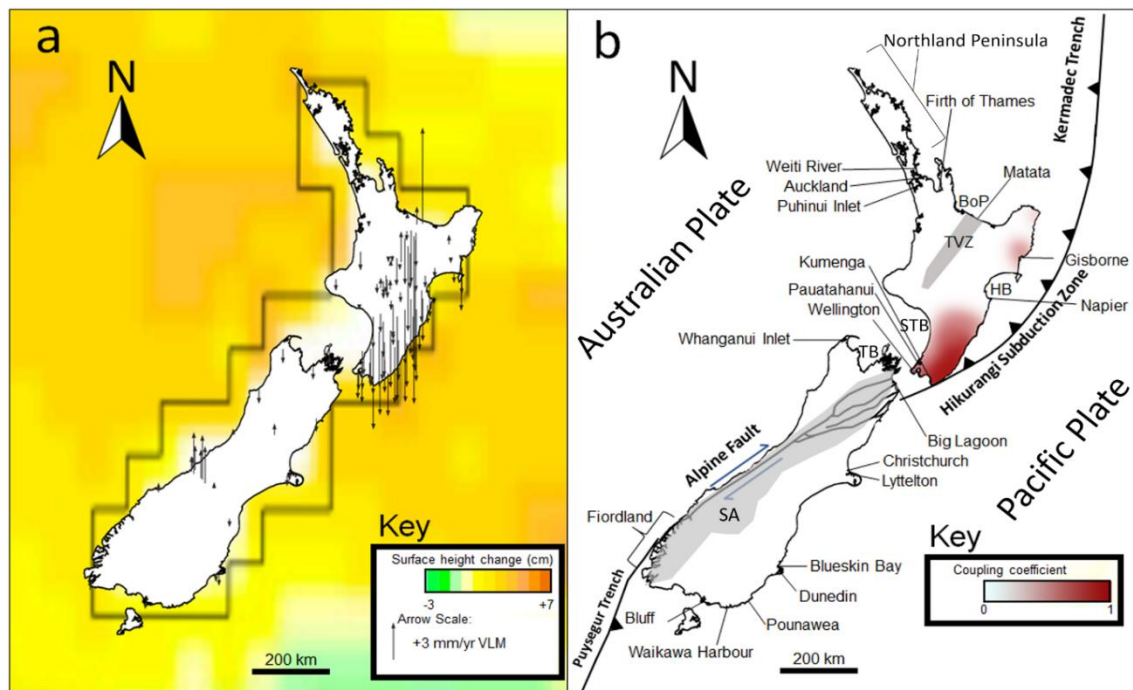
1082 Data Sharing Statement

1083 Data sharing is not applicable to this article as no new data were created or  
1084 analyzed in this study.

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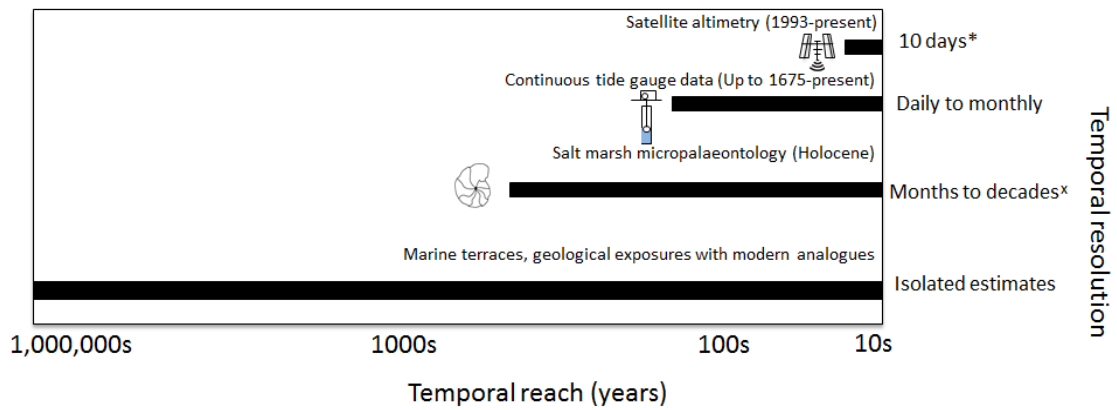


1086 Figure 1. Maps displaying a) the trends of VLM in New Zealand as given by GeoNet's  
 1087 cGPS sites, relative to ITRF2008 (Tenzer and Fadil 2016), and the mean absolute sea-  
 1088 surface height change rise predicted by IPCC's AR4 AOGCMs under emissions  
 1089 scenario A1B (which assumes a balanced emphasis on all energy sources), for the  
 1090 interval 2080-2099, relative to the mean sea-surface height over the interval 1980-1999  
 1091 (Ackerley et al. 2013). b) The 15-year mean degree of coupling between the Australian  
 1092 and subducting Pacific Plates (after Wallace and Beavan (2010)), note that subsidence is  
 1093 highest where coupling is greatest. All locations named in the paper are labelled for  
 1094 context. TVZ = Taupo Volcanic Zone (shaded region), SA = Southern Alps (shaded  
 1095 region) BoP = Bay of Plenty, HB = Hawke's Bay, STB = South Taranaki Bight, TB =  
 1096 Tasman Bay.  
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1098  
 1099  
 1100 Figure 2. A summary of the methods of generating sea-level records in New Zealand,  
 1101 their temporal resolution and the time intervals for which they can be used. Tide-gauge  
 1102 data exists in New Zealand since at least 1900 (Hannah and Bell 2012). \*The ten-day  
 1103 resolution listed for satellite altimetry is the resolution given by Ablain et al. (2019).  
 1104 \*The resolution given for salt-marsh micropalaeontology is dependent upon  
 1105 sedimentation rates, accuracy of dating techniques, and the differing speed of response

1106 to environmental change by different microfossil groups.

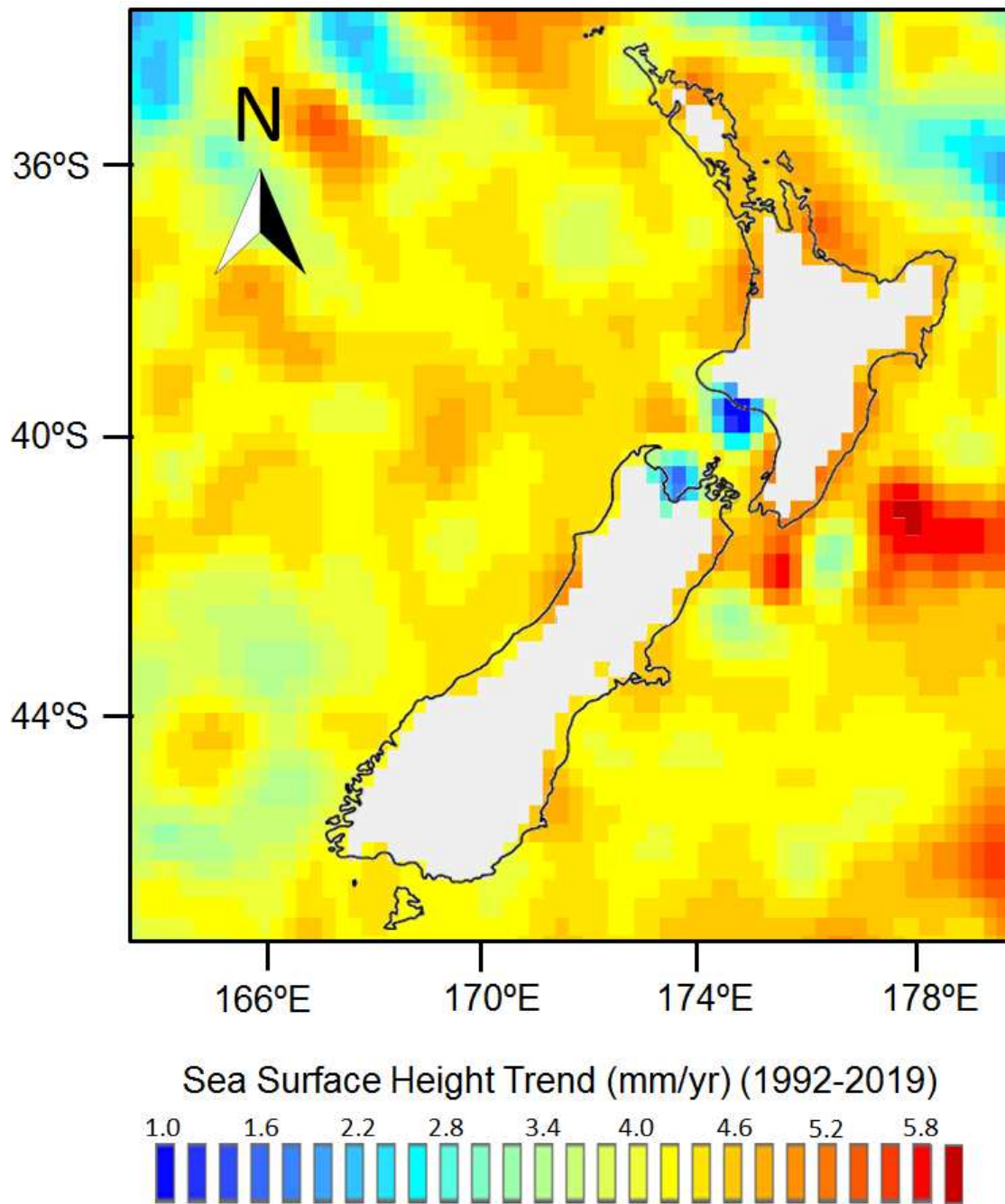


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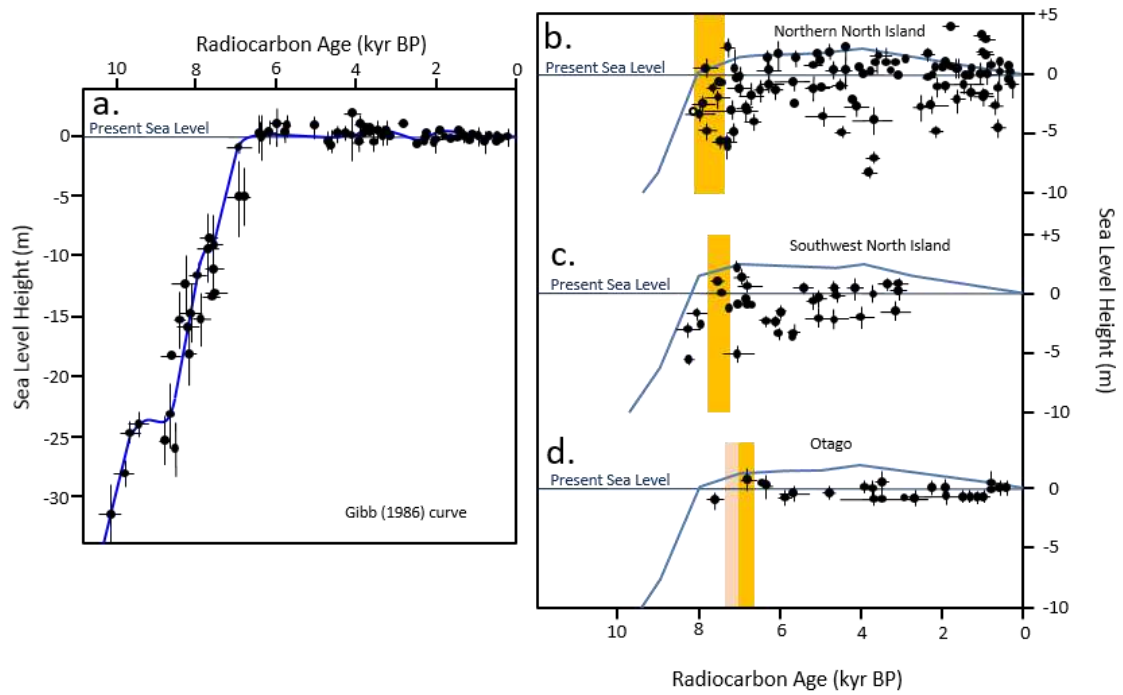
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1110 Figure 3 Map displaying the mean change in sea-surface height, relative to the Earth's  
1111 geoid, around New Zealand September 1992 and May 2019. Data is averaged between  
1112 all Topex/Poseidon and Jason 1 and 2 satellites during this interval, accessed via  
1113 AVISO (2019).



1114

1115 Figure 4. Holocene relative sea-level curves for New Zealand, as determined by Gibb  
 1116 (1986) (a), and three of the regional curves by Clement et al. (2016) (b-d). The locations  
 1117 and types of proxies used are detailed in the respective studies. As discussed in the text,  
 1118 the regional curves display an increasingly later onset of the first attainment of present  
 1119 mean sea-level (shaded) southward. The curves displayed on b-d reflect GIA-modelled  
 1120 predictions of relative sea-level, as discussed in the text of Clement et al. (2016).  
 1121 Modified after Gibb (1986) and Clement et al. (2016).

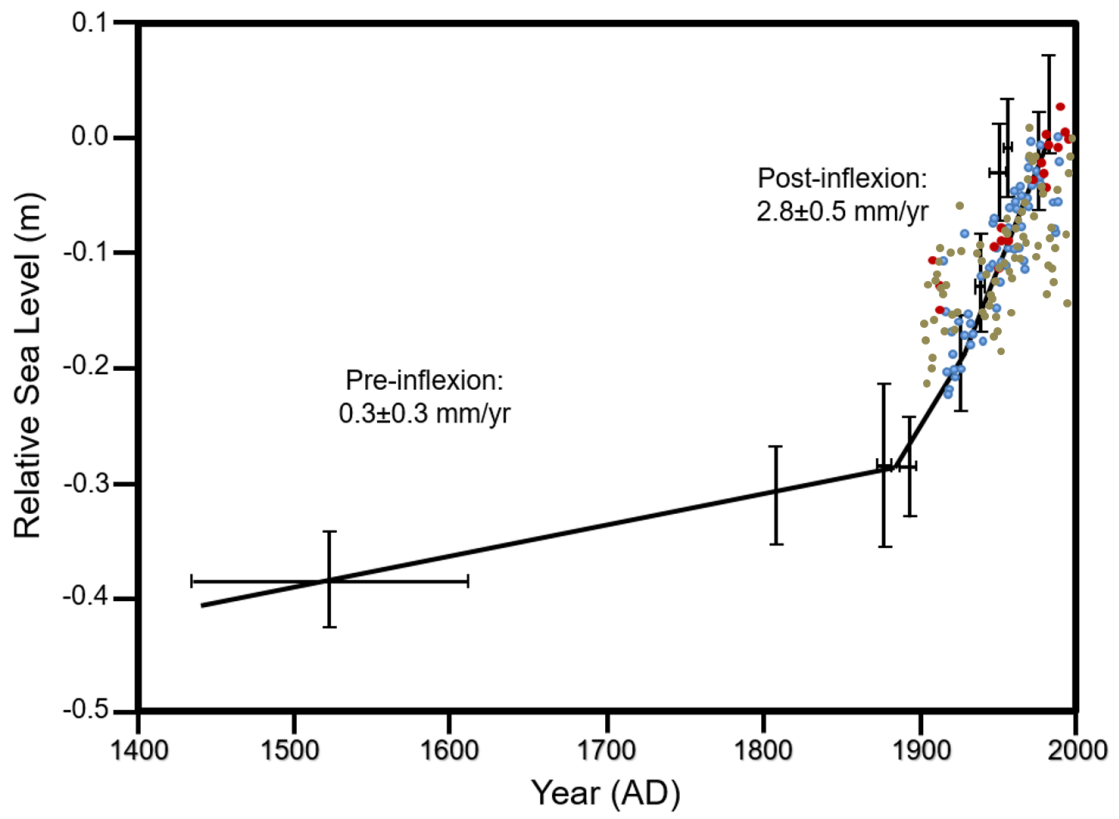


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1123

1124 Figure 5. The relative sea-level curve derived from the Pounawea salt marsh by Gehrels  
 1125 et al. (2008), plotted with the annual sea-level data recorded at the Lyttelton (blue dots),  
 1126 Bluff (red dots), and Dunedin (green dots) tide gauges. Modified after Gehrels et al.  
 1127 (2008), with additional data from the Permanent Service for Mean Sea Level (Holgate  
 1128 et al., 2013).

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