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# Biostratigraphic evidence of dramatic Holocene uplift of Robinson Crusoe Island, Juan Fernández Ridge, SE Pacific Ocean

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## Abstract

A study of the biostratigraphy and sedimentology of Holocene deposits on Robinson Crusoe Island (RCI) on the Juan Fernández Ridge (JFR) indicates that a dramatic but localized uplift occurred since 8000 BP, at a rate of about  $8.5 \text{ mm yr}^{-1}$ . In fact, 5 supratidal flats and sand layers with marine gastropods (mostly *Nerita* sp.) are now exposed ca. 70 m a.s.l., and covered by transitional dunes. The last volcanic activity on RCI occurred at ca. 0.8 Ma (active hotspot located 280 km further west) and there is no sign of a compensating bulge that explains this uplift, isobaths of the sea floor instead suggesting general subsidence. However, modeling indicates that large-scale 10 landslides followed by isostatic rebound are a viable explanation, partially reflected in the low-resolution bathymetry of the area.

## 1 Introduction

The Juan Fernández Ridge (JFR), located on the Nazca Plate in the Pacific Ocean off central Chile (Fig. 1), is a 800 km long seamount and volcanic island chain extending 15 E–W at latitude  $33^\circ \text{S}$ . It has been interpreted as the expression of a fixed hotspot (Von Heune et al., 1997; Montelli et al., 2006) related to a primary mantle plume in the sense of Courtillot et al. (2003) or as part of a “hot line” (Bonatti et al., 1977). According to the plume hypothesis, active diapiric plumes arising from the core-mantle boundary form submarine volcanic *plateaux* or island chains within oceanic plates. As 20 the plate moves away due to sea-floor spreading, these volcanoes are extinguished and new volcanic edifices arise over the active hotspot, forming island chains such as the Hawaiian-Emperor seamount chain. The plume hypothesis has been challenged, however, because some of its predictions have not been confirmed by observation, and plate tectonic processes are thought to be playing a role (e.g., Foulger, 2010).

25 A remarkable feature of hotspot ocean islands is their complex history of vertical displacement (e.g., Ramalho et al., 2013). Uplift and subsidence, as earlier noted by

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Charles Darwin in the 19th century, mostly respond to the growth of the underlying swell and the related isostatic rebound. However, giant landslides can also trigger sudden uplift as inferred for archetypical hotspot volcanoes such as Hawaii (Smith and Wessel, 2000).

Here we present evidence of a very high Holocene uplift rate on Robinson Crusoe Island (RCI) and discuss a possible mechanism together with implications for the long-term evolution of this ocean island.

## 2 Geological and geomorphological background

The JFR is an island–seamount chain largely formed by the Miocene (ca. 9 Ma) O'Higgins guyot and seamount (Von Heune et al., 1997), with lavas dating back to ca. 4 Ma on RCI and nearby Santa Clara Island, and ca. 1 Ma on Alejandro Selkirk Island about 120 km away (Farley et al., 1993). The relief on the western, arid part of RCI is characterized by coastal cliffs bordering a terrace at about 70 m a.s.l., which is especially well developed in the southwestern panhandle (Fig. 1). This terrace is formed on top of a post-shield volcanic platform of middle Pleistocene age from which pyroclastic cones emerge, reaching a maximum elevation of 915 m (Lara et al., 2013, 2014). Holocene sedimentary deposits are restricted to the terrace in the vicinity of Tierra Blanca Bay. The latter name is applied (in Spanish) to a succession described by Morales (1987) as poorly consolidated, calcareous sandstones at the base grading upward into tuffaceous sandstones with numerous fossils. In the transition zone are *Acanthina* and *Lima* fossils with bryozoa fragments, whereas the tuffaceous sandstones host *Luccinea*, *Distoechia*, *Bythinia*, *Orcula*, *Tropicorbis*, *Ena*, and *Cyrena* spp. indicating a Pleistocene–Holocene age, based on a similar fossil assemblage on the continent at this latitude (Covacevich, 1971; Valenzuela, 1978). The Bahía Tierra Blanca Formation has its base at a variable elevation but generally at ca. 70 m above the present mean sea level, where active, incipient barchan dunes (Morales, 1987), partially rework the succession described above.

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### 3 Material and methods

Field campaigns were carried out on RCI in 2011–2013, during which geological mapping was undertaken, stratigraphic sections were measured, and samples were collected for further analysis. Laboratory work consisted of fossil identification, petrographic microscopy, sieve and Mastersizer 2000 (Malvern Instruments, Malvern, UK) analysis of the sediment grain-size distribution, and radiocarbon dating (Beta Analytic Inc., Miami, Accelerator Mass Spectrometer) of gastropods. The latter were collected mostly from sites 1 and 5 (Fig. 1). Several specimens were hand-picked from bulk samples and three selected for dating based on their stratigraphic position and systematics. AMS radiocarbon dates were first corrected for the global marine reservoir effect (e.g., Ulm, 2006) with the Marine IntCal09 calibration program (Reimer et al., 2009). For the localized reservoir correction a Delta-R value of  $373 \pm 76$  from a site nearby (<http://radiocarbon.pa.qub.ac.uk/marine/>) was used. Elevations were measured with a dGPS Trimble® NetRS® and barometric altimeters with respect to the current sea-level and corrected for the regional sea level and daily variation (<http://www.shoa.cl/mareas/tablademarea.html>) with a nominal uncertainty of 5 m.

### 4 Depositional environments

Four lithostratigraphic units and three lithofacies were identified in the Bahía Tierra Blanca Formation, which reaches a total thickness between 2 and 4 m at any specific locality (Fig. 2).

Unit 1 is largely composed of facies 1, which discordantly overlies weathered, basaltic lavas. It consists of very poorly consolidated, slightly calcareous, reddish brown to reddish purple deposits ranging in size from very fine sandstone to claystone. Their composition is made up of volcanic ash mixed with the underlying, weathered lava material. These deposits contain up to 2 % bioclasts (mostly marine bivalves) together with pellets. The most striking feature of this facies is ubiquitous, up to 1 m diameter

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teepee structures, which display prominent edges elevated 3–5 cm above the central parts (Fig. 2). The cracks have been filled in by sands from the overlying unit. Locally, shallow channels and rill marks are present.

While the reddish to purplish brown color suggests a mainly subaerial environment, the presence of teepee structures with elevated rims indicate frequent flooding and drying cycles. These, as well as the occurrence of pellets, are typical of supratidal flats (e.g., Assereto et al., 1977), which concurs with the presence of shallow channels probably reflecting tidal creeks. The general scarcity of hard-shell fossils in this facies can be interpreted as representing a generally hostile environment subjected to frequent dry periods between spring high tides, followed by seawater flooding that would kill land-dwelling snails and other organisms. Marine shells washed in during spring high tides would probably accumulate along the shoreline. Soft-bodied forms more tolerant to such conditions, on the other hand, would not be preserved in such an oxidizing environment.

Facies 2 is present in units 2 and 3, which differ mainly in the darker brown color of the latter due to a thin brownish film coating the grains. Both units 2 and 3 show large-scale, low angle planar cross-bedding and horizontal lamination, but in unit 3, high-angle planar and trough cross-bedding are locally present (Supplement, Fig. A). The 1–2 cm thick cross-beds are formed by alternating light and darker-colored grains without any evident gradation. Rhizocretions are present in the uppermost parts of both units, where individual forms may reach 1.5 m in length (Fig. 3). Although rhizocretions and vertebrate burrows are generally rare, some parts have a fairly high density of the former. Unit 2 is capped locally by whitish calcrete indicating incipient pedogenesis. Gastropods such as *Succinea*, *Fernandezia*, and *Nerita* occur in the middle to upper part of unit 2. Petrographically, the sandstone is well sorted with subrounded grains, lacking a matrix, and cement being only locally present. Bioclasts compose around 55 % of the rock, including brachiopod and pelecypod fragments, echinoderm spines, bryozoa, red algae, foraminifers, and sub-rounded pellets. The rest of the composition is made up of volcanic fragments and minerals such as K-feldspar, plagioclase,

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clinopyroxene, and olivine, with rare quartz. Grain-size analysis of several samples from this facies shows a small traction load, a prominent and very well-sorted saltation load, and a medium- to well-sorted suspension load (Supplement, Fig. B). This facies is interpreted as reflecting coastal eolian deposits perhaps locally affected by weak wave action. This is supported by the reddish brown color of the sandstones, their predominantly fine grain-size with cumulative curves typical of wind-blown deposits, and the presence of the land-dwelling snails *Succinea* and *Fernandezia*, as well as root and burrow systems. The horizontally laminated strata probably formed in sand sheets between low dunes, which might have been parabolic in shape as suggested by the dominance of low-angle planar cross-bedding. Some were subsequently converted into *dikaka* dunes by vegetation. Some low-angle cross-bedding might represent reworking by dissipated wave action during storms and spring high tides along the landward edges of wide supratidal flats. This could also explain the presence of thick-shelled *Nerita* (a marine species) in unit 2. The presence of fragmented marine invertebrates indicates a marine source for most of these sands, which suggests that they formed at a low elevation a.s.l.

Facies 3 is composed of greyish white, medium sorted sandstones interbedded with gravel. The sandstones consist of bioclasts (45–57 %) mostly represented by marine shell fragments including bivalves, gastropods such as *Succinea*, bryozoa, algae, and foraminifers, together with lithic volcanic fragments (27–45 %) and volcanic minerals such as pyroxene, olivine, and felsic minerals (10–17 %). The gravels are greyish brown and matrix- to clast-supported, with the clasts reaching up to 5 cm in diameter. They are mainly volcanic and angular. Locally, calcretes are present at the top of this facies. This facies clearly represents fluvial deposits, probably consisting of shallow, quick-flowing ephemeral streams with gravelly channels and sandy bars. These most likely drained exposed basalts on the fringes of the eolian sand sheets, but also reworked the latter to incorporate the marine bioclasts.

# 5 Radiocarbon dating

Specimens of *Nerita* (SOM, Fig. C) from the eolian sandstones of unit 2 yielded calibrated radiocarbon ages between 8320 and 8030 BP (conventional radiometric age of 7860 ± 40 years BP). Values corrected for the global marine reservoir effect (with a local Delta-R of 373 ± 76 as obtained for the similar entry at <http://radiocarbon.pa.qub.ac.uk/marine/>) correspond to 7550 ± 90 years BP (see Table 1). These marine species were probably reworked from the supratidal flats of unit 1 and would thus represent the age of the latter. On the other hand, land-dwelling species as *Succinea* and *Fernandezia* (e.g., Odhner, 1922) from units 3 and 4 gave calibrated radiocarbon ages between 5440 and 5090 years BP (4580 ± 30 conventional years BP) and 7680 and 7580 years BP (6790 ± 40 conventional years BP), respectively.

# 6 Discussion

The stratigraphic succession of the Bahía Tierra Blanca Formation suggests that supratidal conditions existed in the southwestern panhandle of RCI between 8320 and 8030 years BP (horizons with *Nerita*). It is unlikely that the tides reached more than 1–2 m above the mean sea level, because topographic tide-enhancing conditions such as funnel-shaped estuaries could not have existed due to the absence of large rivers on this part of the island. These supratidal flats were encroached upon by eolian coastal plain deposits at around 5430 years BP (horizons with *Succinea* and *Fernandezia*) and finally fluvial sedimentation as the sea-level receded further during the late Holocene Climatic Optimum (Davis et al., 2003; Koshkarova and Koshkarov, 2004), when the southwestern panhandle would have received more rain.

The present elevation of the supratidal deposits on a marine terrace at 70 m.a.s.l. indicates a very rapid relative sea-level fall since that time. Furthermore, it can be assumed that the eolian deposits of units 2 and 3 were also not more than a few meters above the tidal flats, as they had apparently been reworked locally by waves.







## 7 Conclusions

Large-scale landslides around ocean islands can probably be attributed to an increase in local slopes generated by the construction of volcanic edifices and the development of rifting. At RCI there has been no major surface volcanic activity since about 3 Ma, with only minor post-shield activity at 0.8 Ma (Lara et al., 2013). Nevertheless, the topography of RCI is even steeper than that of Hawaii, which could have allowed sliding to take place. As modeled by Smith and Wessel (2000), directed giant landslides generate isostatic rebound which is larger over the failed flank and spatially asymmetric. Apparent tilting of the Pliocene volcanic pile could be another expression of this process.

Thus, biostratigraphic evidence of the exposure of former supratidal flats 70 m above the present sea level, could be related to a large Holocene landslide not previously detected. These findings highlight the importance of biological markers in order to better understand the complex evolution of ocean islands.

**The Supplement related to this article is available online at doi:10.5194/bgd-11-13605-2014-supplement.**

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**Table 1.** Radiocarbon dates for gastropods from RCI.

Site	Sample	Lab. Number	Conventional radiocarbon age (yr BP)	C13/C12 ratio	Calibrated age (cal yr BP) $2\sigma$	Reservoir corrected age Delta- $R = 313 \pm 76$	Calibrated age (cal yr BP) $2\sigma$	Material	Elevation m a.s.l.
5	PS-25-1	Beta-326738-F	6790 $\pm$ 40	−10.8	7680–7580	6480 $\pm$ 90	7507–7165	<i>Fernandezia</i>	85.0804
1	PS-25-7	Beta-326739-R	7860 $\pm$ 40	−8.0	8320–8030	7550 $\pm$ 90	8508–8050	<i>Nerita</i>	69.7153
1	PS-25-7	Beta-307410-F	4580 $\pm$ 30	−8.4	5440–5090	4270 $\pm$ 80	4965–4522	<i>Succinea</i>	69.7153

Data obtained at Beta Analytic Inc., Miami, Florida.

Elevation computed from dgps data with correction for daily variation of sea level and local height of the antenna.

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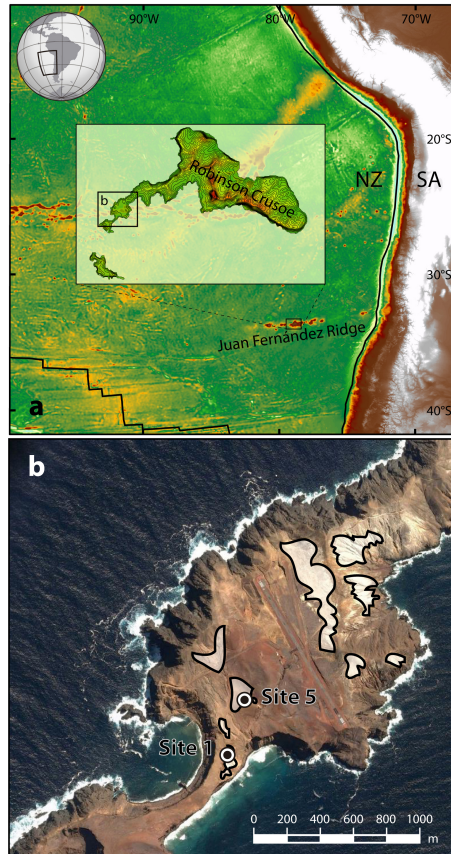
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**Table A1.** Calculation of required wind speed. All equations can be found in Le Roux (2005).

Shell density (calcite)	$\rho_s = 2.85 \text{ g cm}^{-3}$
Shell shape	Ellipsoid, long axis = 1 cm, intermediate axis = 0.75 cm, short axis = 0.35 cm
Nominal diameter	$D_n = \sqrt[3]{(1)(0.75)(0.35)} = 0.64 \text{ cm}$
Water density	$\rho_w = 0.9982 \text{ g cm}^{-3}$
Water dynamic viscosity	$\mu_w = 0.01 \text{ g cm}^{-1} \text{ s}^{-1}$
Air density	$\rho_a = 0.0012 \text{ g cm}^{-3}$
Submerged density of shell in water	$\rho_v = \rho_s - \rho_w = 2.85 - 0.9982 = 1.8518 \text{ g cm}^{-3}$
Acceleration due to gravity	$g = 981 \text{ cm s}^{-2}$
Dimensionless grain size (water)	$D_{ds} = D_n \sqrt[3]{\frac{\rho_g \rho_v}{\mu^2}} = 0.64 \sqrt[3]{\frac{(0.9982)(981)(1.8518)}{(0.01)^2}} = 168.14$
Dimensionless settling velocity of nominal sphere in water	$W_{ds} = \sqrt{2.531 D_{ds} + 160} = \sqrt{(2.531)(168.14) + 160} = 24.2$
Real settling velocity of nominal sphere in water	$W_s = \frac{W_{ds}}{\sqrt[3]{\rho^2 / \mu g \rho_v}} = \frac{24.2}{\sqrt[3]{(0.9982)^2 / (0.01)(981)(1.8518)}} = 63.69 \text{ cm s}^{-1}$
Real settling velocity of ellipsoid	$W_e = -W_s \left\{ 0.572 \left[ 1 - \left( \frac{D_t}{D_i} \right) \right]^{2.5} - 1 \right\} = -63.69 \left\{ 0.572 \left[ 1 - \left( \frac{0.75}{1} \right) \right]^{2.5} - 1 \right\} = 62.55 \text{ cm s}^{-1}$
Dimensionless settling velocity of ellipsoid in water	$W_{de} = W_e \sqrt[3]{\rho^2 / \mu g \rho_v} = 62.55 \sqrt[3]{(0.9982)^2 / (0.01)(981)(1.8518)} = 23.76$
Dimensionless critical shear stress in air for $W_{de} > 11$ , assuming that $\beta_c$ levels off as in water	$\beta_c = 0.00664 \log_{10} W_{de} + 0.00936 = (0.00664)(1.3758) + 0.00936 = 0.0185$
Critical shear velocity $U_{*c}$ in air	$U_{*c} = \sqrt{\frac{\beta_c g D \rho_v}{\rho}} = \sqrt{\frac{(0.0185)(981)(0.64)(2.85 - 0.0012)}{0.0012}} = 166 \text{ cm s}^{-1}$
Assuming a fully rough boundary, required wind speed measured 10 m above the ground	$U_a = U_{*c} \left[ 2.5 \ln \left( \frac{z}{D} \right) + 8.5 \right] = 166 \left[ 2.5 \ln \left( \frac{1000}{0.64} \right) + 8.5 \right] = 4462.9 \text{ cm s}^{-1} \approx 160 \text{ km h}^{-1}$

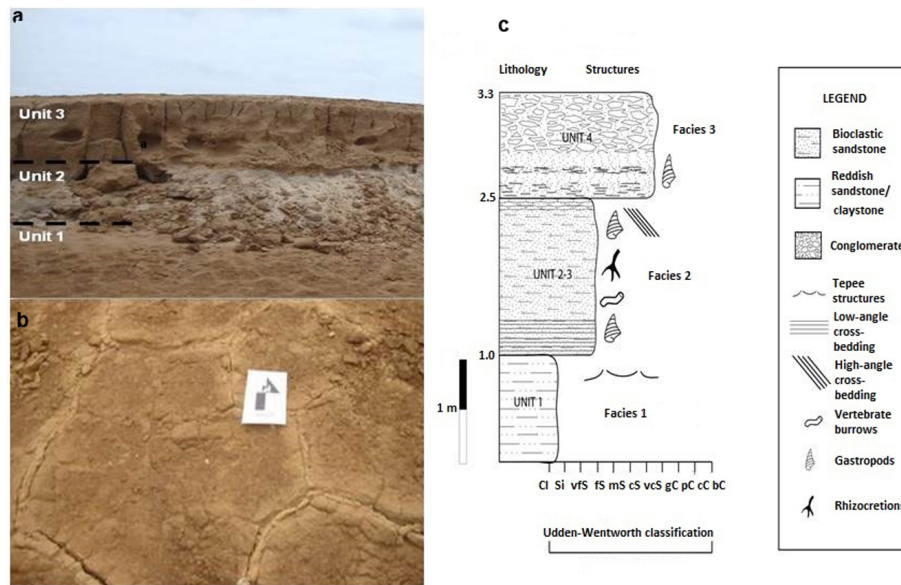


**Figure 1.** Location of Juan Fernández Ridge (a), with Robinson Crusoe Island in a box. Below (b) is a satellite image of the southwestern “panhandle” where the aerodrome is situated. White areas are those of the Bahía Tierra Blanca Formation, where a well exposed supratidal Holocene sequence was dramatically uplifted (see text for details). Sampling sites labeled with numbers (see Table 1). NZ: Nazca Plate: SA: South American Plate.



# Biostratigraphic evidence of dramatic uplift of Robinson Crusoe Island

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**Figure 2.** Exposure of sedimentary units as described in text (a). *Nerita* dated at ca. 8 ka sampled from Unit 1. Below “teepee” structures in Unit 1 (b), interpreted as part of a former supratidal flat. A composite stratigraphic column (c) from records at sites shown in Fig. 1.

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**Figure 3.** *Nerita* shells found in eolian deposits of Unit 2. These are marine species, probably incorporated into dunes developed close to the supratidal flat shoreline. Visual field is 2.5 cm.

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