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Connecting climate and human settlement in the tropical South Pacific: Insights from a lake sediment archive from the Southern Cook Islands

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ABSTRACT

The precise timing and sequence of human arrival into the tropical South Pacific islands is contested and there is ongoing debate around the drivers of migration over the past three thousand years. Recent evidence supports the role of changing climate through palaeoclimate evidence that suggests that the South Pacific has experienced shifts between dry and wet periods throughout the human occupation of the region. Here we focus on the importance of relative drought, as islands are space and resource limited with human populations that depend upon agricultural crops that are rainfall dependent. As such, the occurrence of significant drought events likely threatened early Pacific societies, and such an impact proffers a potential driver of migration. Using lake sediment cores from the island of Mangaia in the Southern Cook Islands, this study utilises stable isotopes, geochemistry and diatoms to create a hydroclimate record extending back 2500 years. We show that prior to the arrival of Polynesians into the Southern Cook Islands there was a significant dry period dating to approximately 885-1075 CE and a further dry period from 1320 to 1460 CE coincident with human settlement of the Southern Cook Islands coinciding with agricultural intensification.

1. Introduction

The Pacific Ocean is home to over 30,000 islands and nearly all of them are in the South Pacific Ocean (United Nations, 1983; Burns, 2002). Today they are inhabited by approximately 12.3 million people (Kirch, 1997; Secretariat of the Pacific Community 2020) . Humans first started to migrate into remote Oceania approximately 1050 BCE, moving through the Solomon Islands, Vanuatu, New Caledonia, Fiji and Tonga before reaching Samoa (Anderson and O'Connor, 2008; Rieth and Cochrane, 2018). Following a 'long pause' of around 1700 years (Thomas, 2008; Hunt and Lipo, 2017), there was a second wave of migration that started in the Cook Islands (Sear et al., 2020) and reached out to the three corners of the Polynesian triangle of Hawaii, Easter Island and New Zealand () (Fig. 1). To explain the timing and sequence of human colonisation of the Pacific, different ideas have been put forward as potential drivers, such as development of navigational skills and canoe technology (Irwin et al., 2022), cultural development (Cochrane, 2018b) and changing climate (Anderson et al., 2006; Sear et al., 2020). Climate - specifically drought - may have put pressure on the resources of certain islands, particularly as populations grew, and this may have forced the Polynesians to migrate eastwards (Anderson, 2002; Goodwin et al., 2014; Nunn and Kumar, 2017; Sear et al., 2020; Hipkiss et al., 2025). Additionally, climate has also been connected to periods of social change (Nunn, 2000; Nunn and Britton, 2001; Allen, 2006, 2010; Field and Lape, 2010; Goff and Nunn, 2016; Rull et al., 2018; Harris et al., 2020). However, climate-human interactions have yet to be properly resolved in the Pacific region, although some modelling work has attempted to examine some of these relationships (Hipkiss et al., 2025).

South Pacific climate is typically characterised by high levels of rainfall that in some parts can exceed 5 m annually, however, the South Pacific islands are also the location of some of the most severe water shortage problems on the globe (White et al., 2007). Annual variability in the climate of the tropical Pacific is characterised by wet and dry seasons with temperature remaining relatively constant year-round. Rainfall in the tropical South Pacific is primarily controlled by two convective bands of rain clouds, the equatorially located Intertropical Convergence Zone (ITCZ) and the South Pacific Convergence Zone (SPCZ), an offshoot of the ITCZ. The location of the ITCZ is typically just

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north of the equator but it shifts northward and southward intra-annually in response to solar insolation (Byrne et al., 2018). The SPCZ runs diagonally across the South Pacific, stretching c. 8000 km from Papua New Guinea, southeast to French Polynesia (Brown et al., 2020). The SPCZ is primarily controlled by the pattern of strong zonal sea surface temperatures as well as atmosphere-ocean interactions including the refraction of Rossby waves, tropical convection from the maritime continent and the strength of trade winds out of the

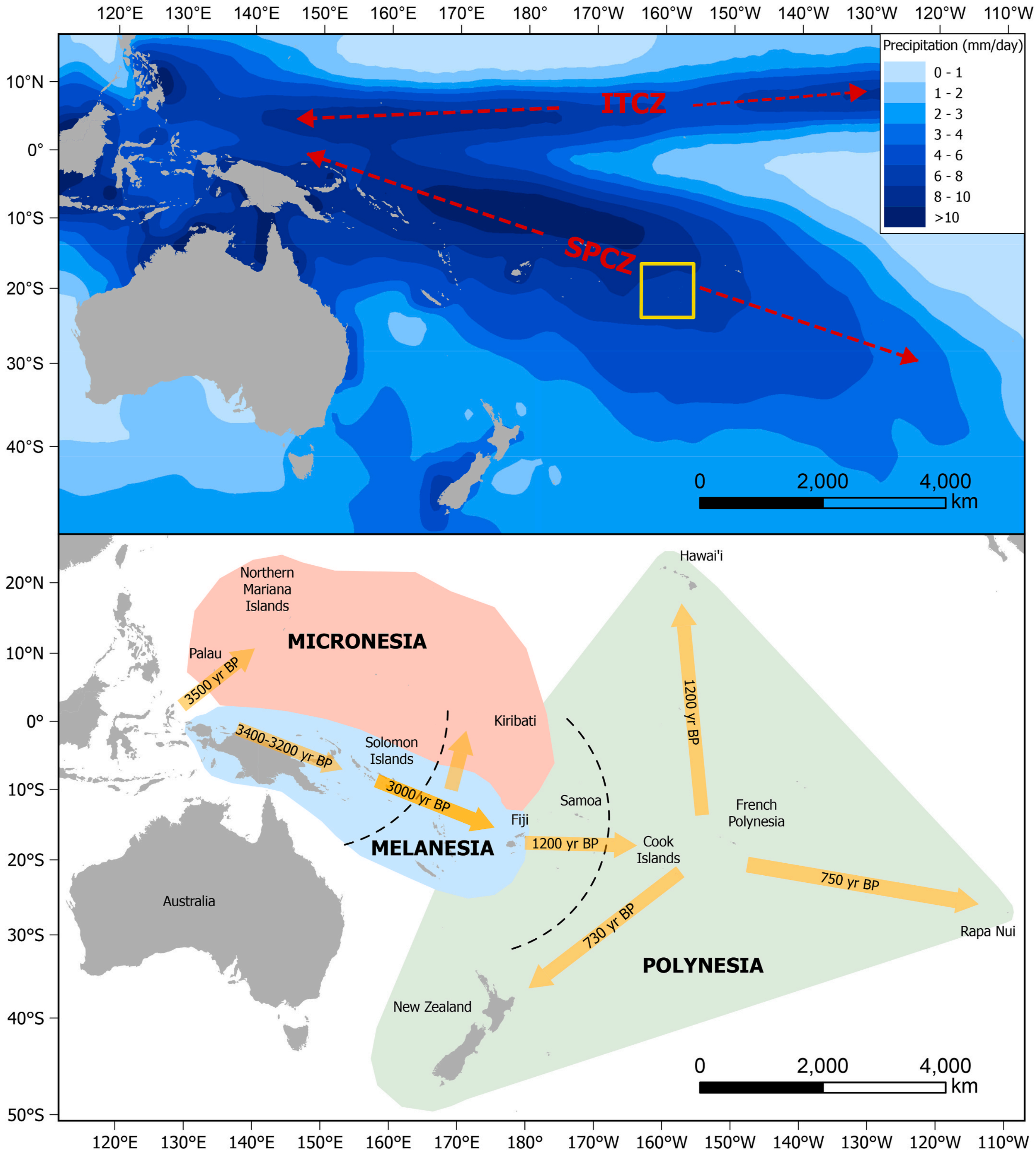

Fig. 1. A: Map of average annual precipitation in the tropical South Pacific, with the location of the ITCZ and SPCZ in relation to the Southern Cook Islands (yellow box). Global Precipitation Climatology Project (GPCP) Monthly Analysis Product data provided by the NOAA PSL, Boulder, Colorado, USA, from their website at https://psl.noaa.gov. B: Map identifying the timing of human arrival to different cultural regions of the Pacific. Dotted black lines denote migration pauses. Updated and based on Matisoo-Smith (2015). Shapefile Data Source: ESRI - ArcWorld Supplement. (For interpretation of the references to colour in this figure legend, the reader is referred to the Web version of this article.)

southeastern Pacific dry zone (Widlansky et al., 2012; Brown et al., 2020). However, hydroclimate in the Pacific is also regulated by a number of modes of climate variability across a range of timescales. These include the El Niño Southern Oscillation (ENSO) and the Interdecadal Pacific oscillation (IPO) that affect sea surface temperatures and atmospheric pressure which in turn impacts upon wind and air temperatures over interannual and interdecadal timescales respectively (Widlansky et al., 2012; McGree et al., 2016). These climatic oscillations also influence the location, extent and intensity of both the ITCZ and SPCZ and therefore hydroclimate variability across island groups in the South Pacific region (Lorrey et al., 2012; Haffke and Magnusdottir, 2013; McGree et al., 2019; Brown et al., 2020; Higgins et al., 2020). During positive IPO phases and El Niño events, the SPCZ location can be altered by as much as 3-5° of latitude and 10° of longitude toward the northeast, though these two oscillations operate quasi-independently (Folland et al., 2002; Brown et al., 2020) so can compound to see greater changes in the SPCZ extent and location (Salinger et al., 2014). For example, during positive IPO phases, which exhibit El Niño like patterns, the SPCZ migrates northeast toward Samoa, and if an El Niño event occurs during a positive phase the SPCZ can shift even further northeast leading to drier conditions for islands in the Southwest Pacific and the same is true of the reverse with southwest movement toward Fiji during negative IPO phases and La Niña events leading to wetter conditions in the southwest Pacific (Widlansky et al., 2012; Salinger et al., 2014). Furthermore, during extreme El Niño events, such as the 1997/98 season, the SPCZ can shift to sit in a near-parallel position in relation to the ITCZ bringing low levels of rainfall and drought conditions to the South West Pacific including the Cook Islands (Widlansky et al., 2012; Haffke and Magnusdottir, 2013; Power et al., 2014; Salinger et al., 2014). Periods of both drought and flood are prominent features in the South Pacific climate due to changes in ENSO events and IPO phases, which are strongly related to precipitation anomalies in the southwest Pacific (McGree et al., 2016). These changes in the SPCZ location and extent alter the amount of rainfall falling on different island groups and impact on the people living there who are dependent on rainfall for drinking water and agriculture (Brown et al., 2020; Faraji et al., 2022).

The South Pacific is a region we know relatively little about in terms of climate variability on decadal to millennial timescales despite the region's importance in influencing global climate (Brown et al., 2020). Palaeo records also suggest that shifts in precipitation were larger and longer than observed in the instrumental record (Maupin et al., 2014; McGree et al., 2016; Sear et al., 2020; Maloney et al., 2022). New records are required to address the lack of understanding of long-term hydroclimate variability, which has hampered efforts to test the drought hypothesis as a driver of migration and settlement in the South Pacific. Archives of longer-term palaeoclimate can be found in sediment deposits that have accumulated at relatively high rates throughout the Holocene, including lakes and swamps. Recent evidence from lake sediment archives in the Pacific has demonstrated the potential to provide long (c.7000+ yr records) (Gosling et al., 2020; Sear et al., 2020) and higher resolution (50-150 yr resolution) hydroclimate records from the Pacific islands (Atwood and Sachs, 2014; Sear et al., 2020). Here we reconstruct climate over the past 2500 years with the aim to tie together climate and archaeological records to understand whether climate, and specifically drought, played a role in the timing of human migration and subsequent settlement of islands in the tropical South Pacific. We will present a new palaeoenvironmental record from Lake Tiriara in the Cook Islands, which represents a key archaeological location in relation to the second wave of human migration into the Pacific. The lake is also located locally to a key archaeological site – the Tangatatau Rockshelter (Kirch, 2017) – that covers the period of human occupation allowing potential connections to be drawn between the palaeo-records developed at Lake Tiriara to the archaeological record in the region. Mangaia also sits in an SPCZ sensitive area whereby changes in its location can cause a shift in the amount of precipitation the island receives making it vulnerable to changing climate. We utilise a multi-proxy approach to build a new hydroclimate record in the tropical South Pacific over a key period in Pacific human history that includes carbonate oxygen isotopes to reconstruct hydroclimate and μXRF, sedimentology, stable isotopes from organic content and diatoms to identify local changes in the environment over this period.

2. Regional setting

The Cook Islands are a South Pacific island nation made up of 15 islands that are geographically split into a northern and southern group and have a recorded total population of 17,434 (Cook Islands Statistics Office, 2018). The islands are located more than 5000 km from Australia and stretch across two million square kilometres between American Samoa and French Polynesia. Mangaia is the southernmost island of the Cook Islands (21°55′26.7″S 157°55′19.4″W; Fig. 1).

The Cook Islands have a well-defined wet (December-May) and dry (June-November) season with mean temperatures relatively constant for the year, only ranging between 25 and 28 °C. Between 1941 and 1995 rainfall for Mangaia ranged between 1091 and 2717 mm/year with average annual rainfall was 1976 mm/year (KNMI, 2022). During the wet season, the SPCZ shifts towards the southwest to sit over the Southern Cook Islands bringing changeable weather and heavy rain. Over the dry season, the SPCZ shifts northwards away from the Southern Cook Islands and the climate is primarily controlled by the dry south-easterly trade winds (Faraji et al., 2022). The Cook Islands are located in a key position in relation to the SPCZ as they sit just beyond the typical average extent of the SPCZ (Fig. 1A). This means that these islands are particularly sensitive to SPCZ movements and its longer-term changes in location and extent in response to interannual to interdecadal climatic oscillations.

Mangaia is located in the southern Cook Islands group (Nunn et al., 2016). In terms of the island morphology, Mangaia has an inner eroded volcanic basaltic cone that rises to a maximum elevation of 168 m (Ellison, 1994b) and is estimated to be approximately 17-19 million years old (Dalrymple et al., 1975). Around the inner volcanic cone is an encircling outer makatea zone, which is composed from an ancient raised limestone reef (Wood, 1978). The makatea rises as a sharp escarpment, which reaches a maximum height of 70 m and is 700 to 2000 m wide. Between these two zones lies lowland swamps that develop at the edge of the radially draining volcanic cone where water becomes trapped by the makatea cliffs forming wetlands and lakes (Ellison, 1994b; Parkes, 1994) (Fig. 2). On Mangaia today, the central part of the island is dominated by introduced species which are remnants of the transformation of this landscape for agricultural uses. whilst vegetation on the makatea is dominated by native species (Merlin, 1991).

2.1. Colonisation of the Southern Cook Islands

The Southern Cook Islands have been described as “gateways” for the final phase of human migration into Eastern Polynesia (Allen and Wallace, 2007) and so the timing of human arrival on these islands has been widely debated. Wilmshurst et al. (2011) suggested that human settlement of the southern Cook Islands occurred around 1250 – 1281 CE based on a meta-analysis of radiocarbon dates from archaeological sites. For Mangaia, early work by Ellison (1994b) and Kirch and Ellison (1994) put the arrival of humans on the island at c. 550 BCE as evidenced by a peak in charcoal. Work by Chagué-Goff et al. (2016) narrowed the window of human arrival to 67 BCE – 398 CE These dates however are contentious, and a recent re-analysis of archaeological evidence puts the arrival and initial settlement of Mangaia at 990-1180 CE (Kirch, 2017). This is corroborated by recent evidence from the island of Atiu also located within the Southern Cook Islands (Sear et al., 2020).

For the purposes of this study, human arrival for Mangaia is set at 1000 CE as suggested by Kirch (2017). The pre-contact human history of

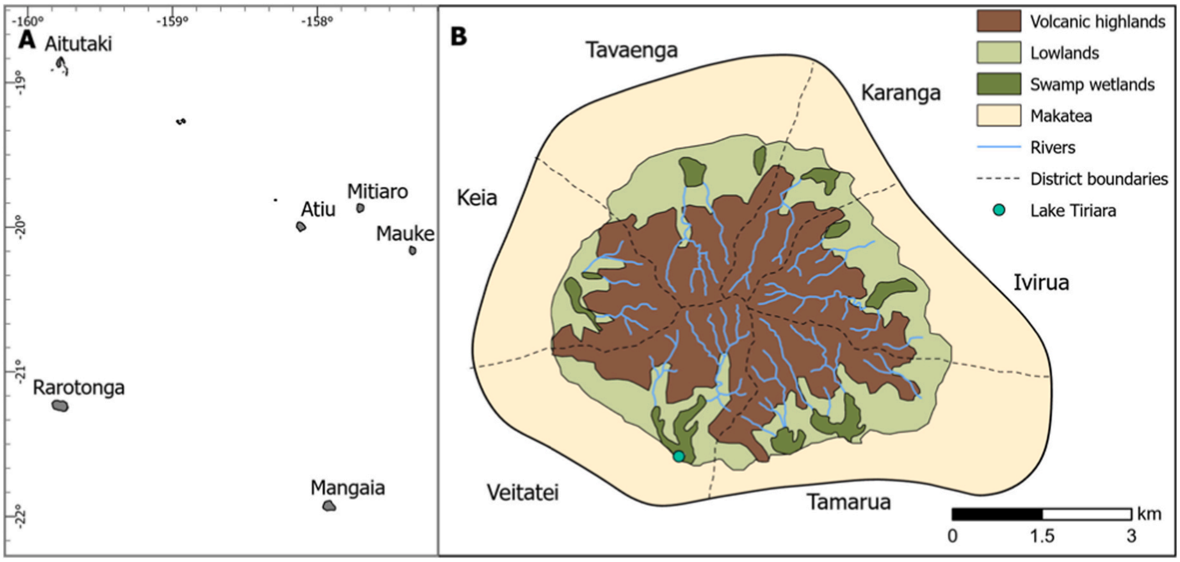

Fig. 2. A: Map of the Southern Cook Islands. Shapefile Data Source: ESRI - ArcWorld Supplement. B: Map of the island of Mangaia showing the major land types, district boundaries, rivers and the location of Lake Tiriara. Based on Kirch (2017).

Mangaia has been split into four phases Vairorongo (1000-1300 CE), Tangatatau (1300-1400 CE), Ngaaitutaki (1400-1600 CE) and Tautua (1600-1830 CE) - based on a compilation of the archaeological research by Kirch (2017). The prehistory period ends with European arrival, initially with contact from the HMS' Resolution and Discovery during Cook's third voyage of the Pacific in 1777 but was consolidated later with the arrival of a permanent missionary presence on the island in 1824 (Williams, 1842).

2.2. Site description

Lake Tiriara (21°57′03.9″S 157°55′45.7″W) is a small freshwater lake located in the Veitatei drainage basin between the volcanic cone and makatea boundary on the south side of Mangaia and is now the only lake on the island likely due to its low elevation and Ellison (1994b) reported that it was filling in rapidly. It has an area of approximately 0.2 km^2 and has a maximum depth of 1.2 m as recorded during coring in 2016. Lake temperature was reported to be 27.6 °C with a pH of 7.37 and conductivity of 1693 μS cm^{-1} as stated by Schabetsberger et al. (2009). Tiriara is thought to be connected to the ocean via 700 m long tunnels through the makatea (Ellison, 1994a; Chagué-Goff et al., 2016).

3. Materials and methods

The Mangaia lake sediments were cored from the deepest parts of the lake during a field campaign in July 2016 using a GeoCore piston system, with cores sealed and kept in cold storage. A UWITEC gravity-type corer was used to extract the youngest sediments and maintain the water-sediment interface. The cores extend down to a maximum sediment depth of 840 cm and are mainly composed of gyttja, silty/clay lake sediments. Here we utilise the top-most 560 cm of sediment to focus analysis on the period just prior to and including human occupation of Mangaia around 1000 CE

3.1. Chronology

Ten AMS ^{14}C dates were produced for the Tiriara sequence using a combination of bulk sediment and picked plant macrofossil samples. Bulk sediment was used in place of plant macrofossils when there was insufficient macrofossil material available. It was not possible to identify whether the macrofossils were terrestrial or not due to the small size of the material. The stable carbon isotope ($^{13}C/^{12}C$) ratio was measured for the smaller sample AMS dates during analysis and that value is used to correct for isotopic fractionation as per Stuiver and Polach (1977), but $\delta^{13}C$ value for the small AMS samples were not reported. Radiocarbon ages are reported as the number of years before present with present referring to the year 1950 CE. All age models were generated using Bayesian age-depth modelling via the Rbacon R package (Blaauw et al., 2025). The ^{14}C ages were calibrated using the Southern Hemisphere SHCal20 curve (Table 1).

An attempt to date the top of the Tiriara sequence using ^{210}Pb found that data from the top 30 cm of the Tiriara sediment sequence are, within measurement uncertainties, uniform (Fig. S2). This indicates that the sediments from at least 30 cm depth up to the top of the core have been subject to mixing. The ^{137}Cs levels for Lake Tiriara sediments were below detection.

3.2. μXRF

Prior to any sub-sampling, the sediment cores for Lake Tiriara were extruded, split and their surface cleaned before being scanned using an Itrax XRF core scanner (Cox Analytical Systems, Gothenburg) and subjected to magnetic susceptibility analysis. For the μXRF data, where individual elements are presented individually rather than ratios, the element is normalised against the incoherent Compton/Raleigh scatter (inc) to account for water and organic content (Boyle et al., 2015; Gosling et al., 2020). For the PCA, the μXRF data was transformed using a centred log ratio to normalise the data and account for non-linear changes in the matrix and dilution effects (Weltje et al., 2015; Bertrand et al., 2024).

3.3. Magnetic susceptibility

Magnetic susceptibility was used as a measure of erosional activity in the catchment and associated inwash (Dearing, 1999b; Sear et al., 2020). The magnetic susceptibility of the Lake Tiriara cores was measured using the Bartington Instruments MS2K at 1 cm resolution. To correct and account for the background level of magnetism, each of the individual measurements of the sediment core were taken between a pair of blank air measurements as per Dearing (1999a).

3.4. Stable isotopes, TOC and CO_2 yield

Oxygen and carbon isotope analysis ($^{18}O/^{16}O$, $^{13}C/^{12}C$) of

Table 1
^{14}C dates from Lake Tiriara, Mangaia. NR – Not reported.

Lab ID	Sample depth from surface (cm)	Dated material	$\delta^{13}C$ (VPDB‰)	Conventional radiocarbon age	2-sigma calibration (CE)
SUERC-103417	107-108	Plant Macrofossil	−15.7	151 ± 35	1661 – 1758
SUERC-103418	190-191	Plant Macrofossil	−27.8	309 ± 37	1322 – 1506
UCIAMS-259598	191-192	Bulk Sediment	NR	650 ± 30	1318 – 1501
UCIAMS-259599	221-222	Bulk Sediment	NR	890 ± 30	1200 - 1380
UCIAMS-259598	290-291	Bulk Sediment	−28.0	983 ± 37	948 - 1146
SUERC-107298	310-311	Bulk Sediment	−27.9	880 ± 37	852 - 1075
UCIAMS-259600	374-375	Bulk Sediment	NR	1575 ± 30	483 – 671
UCIAMS-259601	451-452	Bulk Sediment	NR	1650 ± 30	82 - 309
SUERC-103427	475-476	Bulk Sediment	−30.6	1999 ± 35	47 BCE – 144 CE
SUERC-103428	544-545	Bulk Sediment	−23.7	2509 ± 37	575 – 230 BCE

carbonates from bulk sediment were used as hydroclimate and environmental proxies respectively. The samples for $\delta^{18}O$ and $\delta^{13}C$ from carbonate ($\delta^{18}O_{carb}$ and $\delta^{13}C_{carb}$) were taken to the British Geological Survey (BGS) for processing and were prepared by bleaching the samples using 5% sodium hypochlorite to remove any organic carbon. The sampling interval frequency was increased from 10 cm to 5 cm within the cores where the approximate location of human arrival sat within the sequence based on previous work at the site (Ellison, 1994b; Temple-Brown, 2018). There was no evidence for other biogenic sources of carbonate (shells/ostracods etc) so it is assumed that the bulk sample carbonates measured are authigenic marl and this is consistent through time. Results are reported as per mil, ‰ VPDB (Vienna Pee Dee Belemnite) for $\delta^{13}C_{carb}$ and $\delta^{18}O_{carb}$ from carbonate. CO_2 yield upon carbonate dissolution was used as a proxy for carbonate abundance and is reported as millibar (mb). Many samples from Mangaia had no or limited carbonate and carbonate abundance is very variable. Organic carbon isotopes ($\delta^{13}C_{org}$) along with carbon and nitrogen concentration (TOC and C/N ratio) were used to identify shifts in lake productivity, erosion in the surrounding landscape and catchment vegetation. The samples were taken at the same resolution as the carbonate samples and were acidified using 5% hydrochloric acid to remove any inorganic carbon. Results are reported as per mil, ‰ VPDB (Vienna Pee Dee Belemnite) for $\delta^{13}C_{org}$ and % for C and N.

3.5. Diatoms

Diatoms were used as a proxy for water salinity. Samples were prepared according to an adapted method described by Battarbee et al. (2001) and reported as relative abundance percentage.

3.6. Charcoal

Charcoal can be used as a proxy for fire activity and anthropogenic land clearance in the surrounding area (Whitlock and Larsen, 2001). For the purposes of this study, only macro-charcoal fragments (>125 μm) were counted as they represent local rather than regional burning (Sear et al., 2020; Strandberg et al., 2023). Charcoal samples were taken at 1 cm resolution at 20 cm intervals.

Further details on methods used in this study are available in the supporting information (SI).

4. Results

The Lake Tiriara sequence contains seven zones based on changes in the proxy data. These zones will be used to compare changes across all proxies presented for the sequence (see supporting information (SI), Table S1).

4.1. Chronology

The maximum mean age of the sequence is 520 BCE (682-296 BCE, 2σ) at 560 cm depth and 100% of the dates overlapped with the age model produced using the Rbacon package (Fig. 3). The age model suggests an average accumulation rate of 0.22 cm/yr though with an increase in the average rate of sedimentation at approximately 306 cm (c. 1000 CE) from 0.17 cm/yr to 0.30 cm/yr.

4.2. μXRF geochemistry

Titanium (Ti) presence is typically used as an erosion proxy in lake sediments as it is washed in from the surrounding catchment (Davies et al., 2015; Sear et al., 2020; Maloney et al., 2022). Ti/inc is variable throughout the core but has increased levels in zones 1 to 2 (c. 20-355 CE), towards the end of zone 3 (c. 680-995 CE) and across zone 4 (c. 1150-1625 CE) (Fig. 4). Peaks in Calcium (Ca) have been associated with carbonate precipitation within lake sediments (Burnett et al., 2011; Davies et al., 2015). The Ca/inc is consistently low with values below 0.05 throughout the record. There are two exceptions to this. The first occurs from c. 40 to 345 CE, starting in zone 1 with calcium peaking in zone 2 with a value of 0.17, which occurs around 340 CE, before steadily decreasing through into zone 3. The second occurs from c. 1180 to 1475 CE spread across zone 4. Peaks in Bromine (Br) have remained relatively steady throughout the sequence though the topmost sediments in zone 5 show a much higher level of Br in comparison to the rest of the sequence, indicating a higher organic content in that section of the core (Davies et al., 2015). Strontium (Sr) shows the main peaks occurring in zones 2 (c. 230-380 CE) and zone 4 (c. 1330-1515 CE). Similar to Ca, Sr is associated with carbonate precipitation but particularly aragonite ($SrCO_3$) (Burn and Palmer, 2014). The Tiriara sequence shows some similarity between Sr peaks with Ca. There are higher levels of S in the older sections of the sequence prior to c. 570 CE but this drops between zone 3 and zone 4 to an overall lower average after c. 995 CE. S can indicate the presence of gypsum ($CaSO_4$) (Burnett et al., 2011; Burn and Palmer, 2014), increased input from the ocean (Davies et al., 2015) or sourced from soil leaching (Davies et al., 2015) and organic content (Moreno et al., 2007). Ca has a moderate positive relationship with Sr with an r-squared value of 0.44 ($n = 500$, $p = < 0.0001$) but a weaker relationship with Ti (Fig. 5). This suggests that carbonate precipitation is authigenic rather than being transported via inwash which is indicated by the presence of Ti (Evans et al., 2019).

A Principal Component Analysis (PCA) provides information on how the composition of the core sections changes throughout the core sequence, showing which variables correlate and which sit independently (Muller et al., 2008). The distance along the PCA axes indicates the weighting of each depth in the first two principal components. Axis 1 explains most of the variance (65.7%) and shows key zones spread mainly along this axis (Fig. 6). Overall, PC1 and 2 combined accounts for 79% of the data variance. Br and S have both been used as indicators of organic content (Moreno et al., 2007; Burn and Palmer, 2014) and have a strong negative correlation to the terrestrial indicators. Br and S track negatively in PC1 whilst Br is loaded positively in PC2 and S is loaded negatively. The biplot shows that sections from the youngest and oldest parts of the Tiriara cores are most strongly associated with these elements but with S associated most strongly with zone 3 and zone 1, 4 and

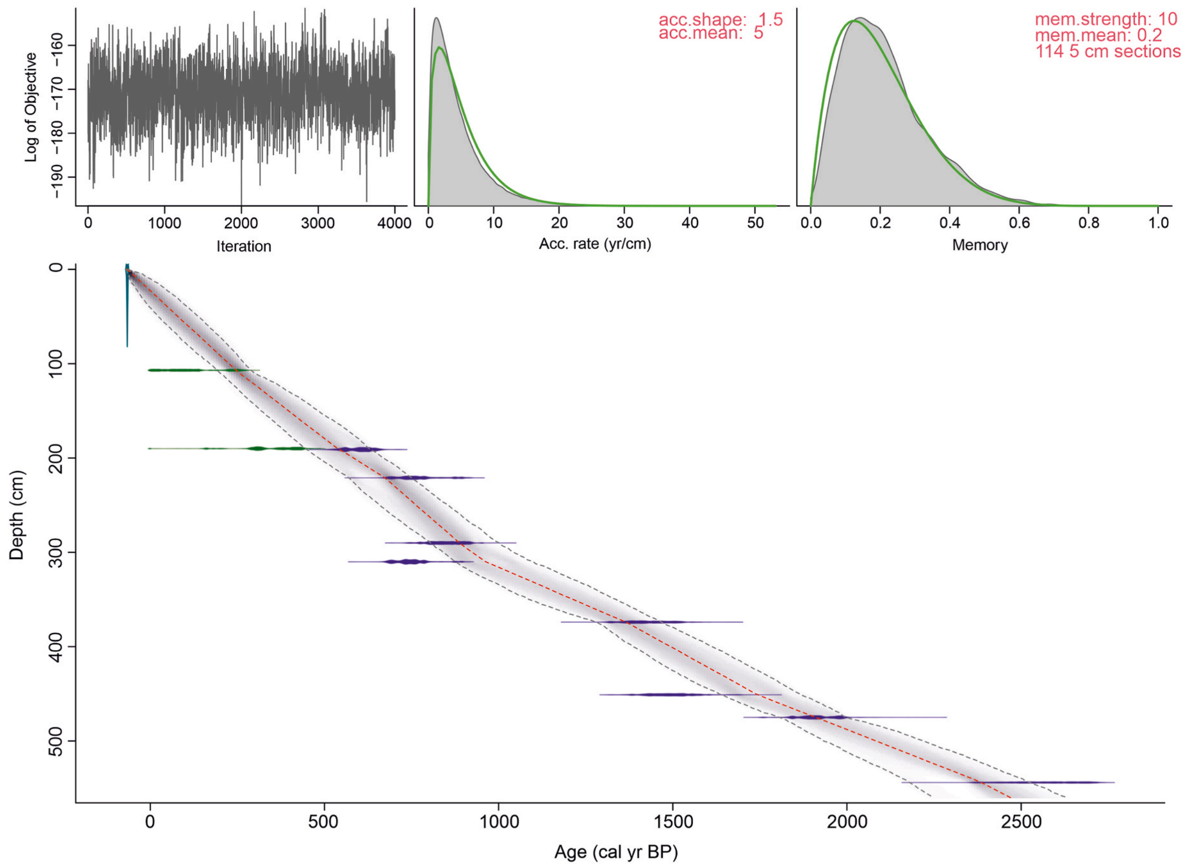

Fig. 3. Age model generated by the Rbacon package for Lake Tiriara showing age (cal yr BP) against depth (cm). This is an unweighted age model and colours denote the material used for each date. Green - Plant macrofossil and Blue - Bulk sediment. (For interpretation of the references to colour in this figure legend, the reader is referred to the Web version of this article.)

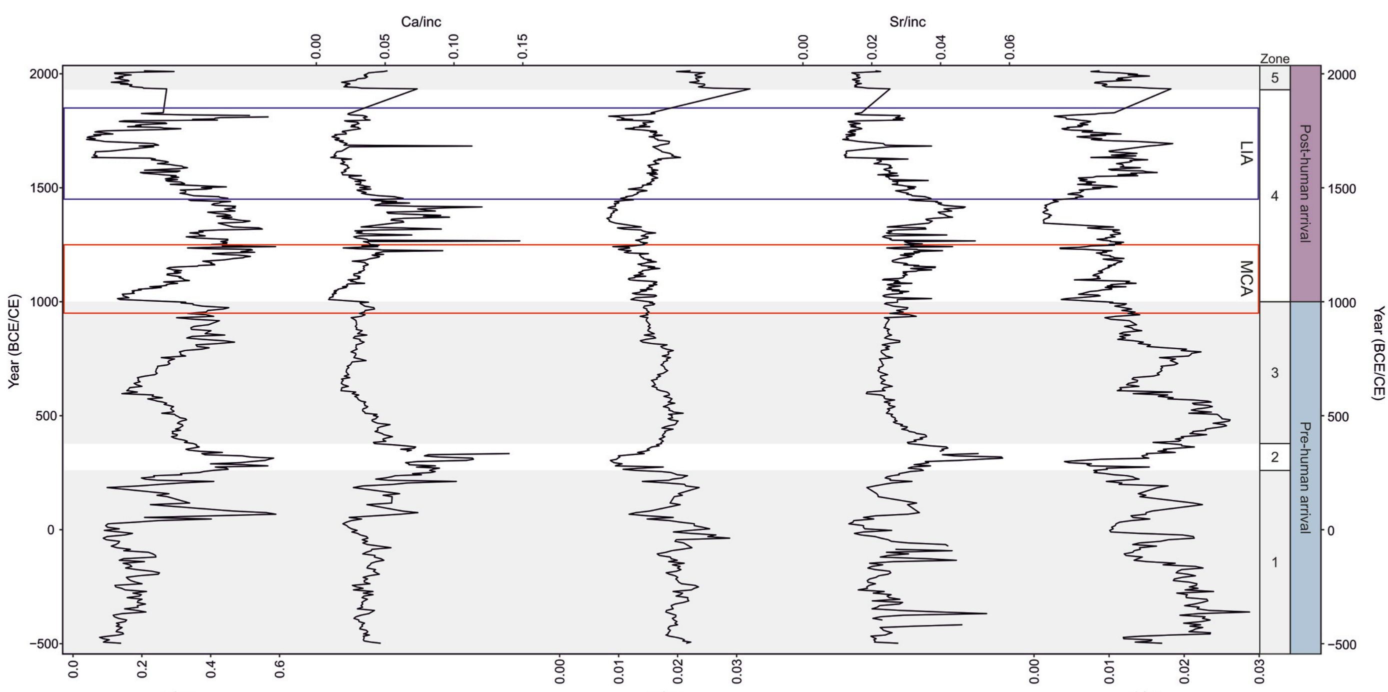

Fig. 4. Plot showing the XRF geochemistry data (Ti, Ca, Br, Sr, S) corrected against incoherent scatter (inc) for the Mangaia sequence. Red box is showing the timing of the Medieval Climate Anomaly (MCA) and the blue box is showing the timing of the Little Ice Age (LIA) (Masson-Delmotte et al., 2014). (For interpretation of the references to colour in this figure legend, the reader is referred to the Web version of this article.)

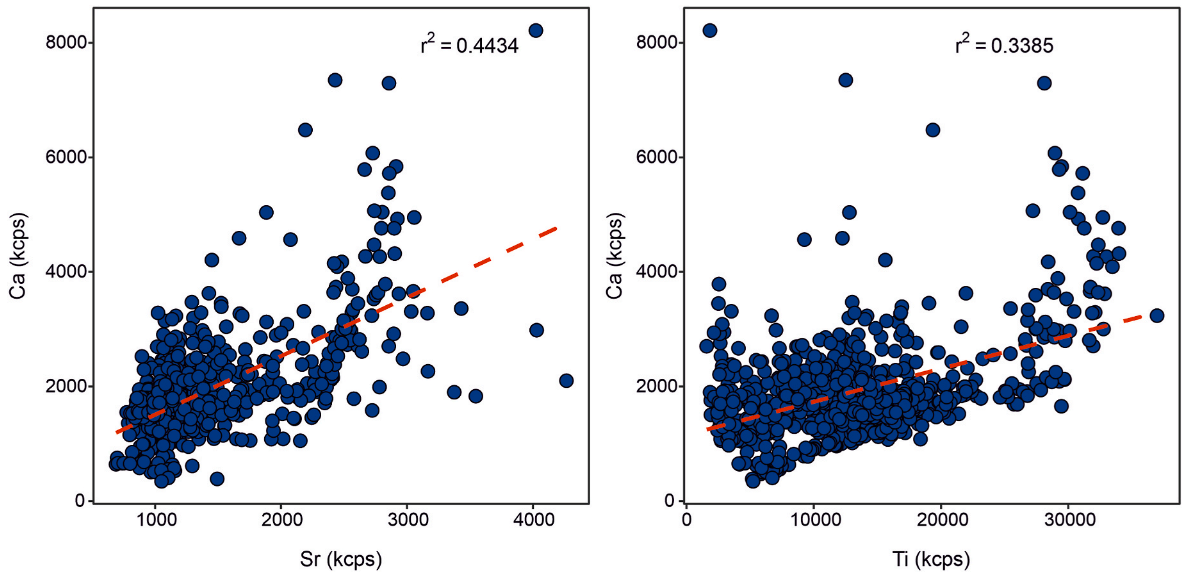

Fig. 5. Plots showing the relationship between Sr against Ca (left) and Ti against Ca (right) in the Tiriara core sequence. Plots show a regression line and display the r-squared value.

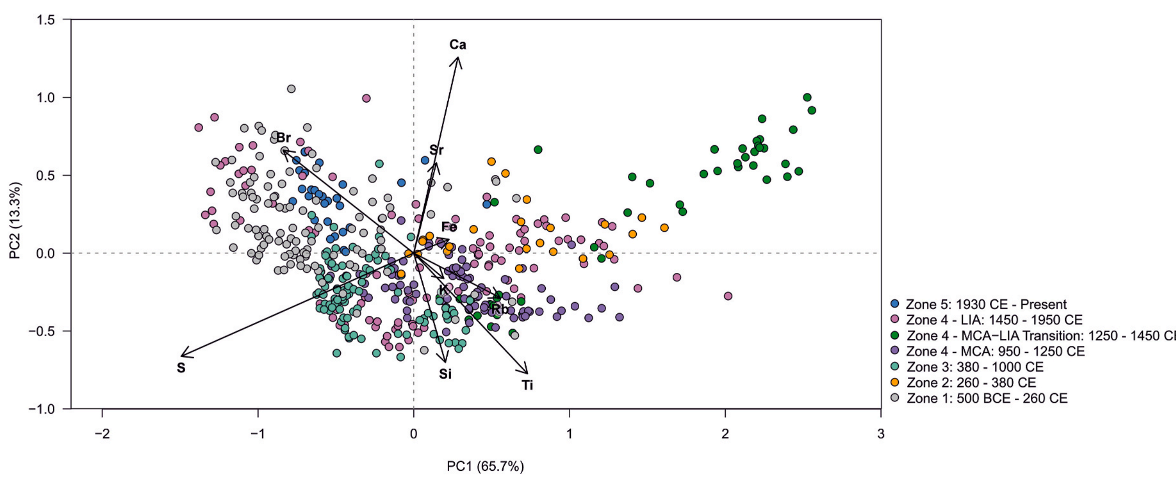

Fig. 6. A principal component analysis plot for PC1 and PC2 for the Lake Tiriara XRF data. Colours denote which zone of the core each point corresponds to as laid out in table SX. (For interpretation of the references to colour in this figure legend, the reader is referred to the Web version of this article.)

5 associated most strongly with Br. The terrestrial indicators (Ti, Fe, K, Rb, Si) generally have a strong positive correlation to one another though this correlation is weak where indicators are positively (Fe) or negatively (Ti, K, Rb, Si) loaded on PC2. These indicators are most strongly associated with zone 4 and 5. Further to this, the freshwater/carbonate indicators do not show a strong correlation with other indicators though Ca shows a moderate correlation of 0.47 with Sr. Whilst Si, which can be either a freshwater or terrestrial indicator, here it appears to have correlation with the terrestrial elements and a weak correlation with Ca of 0.27.

4.3. Magnetic susceptibility

Peaks in magnetic susceptibility are used as evidence for catchment soil erosion and runoff, augmented by changes in sediment availability due to catchment disturbance such as clearance following the arrival of humans. Lower values suggest periods of limited catchment disturbance or dilution by organic matter. There are three large peaks in magnetic susceptibility within the Lake Tiriara record located in zones 2 and 4 (Fig. 7). After all three of these larger peaks, the values return to lower base values which typically happens relatively quickly with a rapid rise and fall in magnetic susceptibility values, but for the second larger peak in zone 4 the values increase relatively quickly but falls more gradually back to lower base values.

4.4. Organic carbon isotopes and C/N abundance

There is a relatively large range for total organic carbon (TOC) in this sequence with a minimum value of 0.9% and a maximum value of 34.5% (Fig. 7). Zone 1 shows the highest TOC values for the Mangaian sequence with maximum values of 34.5%. However, the TOC declines in zone 2 to 0.9% at around 320 CE - the lowest values measured in the sequence. The values through the rest of the sequence are lower on average than zone 1, but range between 1.5 and 18.0% from zone 3 through to the top of the sequence.

The C/N values show a relatively small range from a minimum value of 8.34 at around 320 CE to a maximum of 17.82 at around 285 BCE. The C/N record starts with relatively high values (11-18) across zone 1 (Fig. 7). This event is relatively short-lived, and C/N values rise and vary between 10.5 and 15.8 from zone 3 through to the top of the core

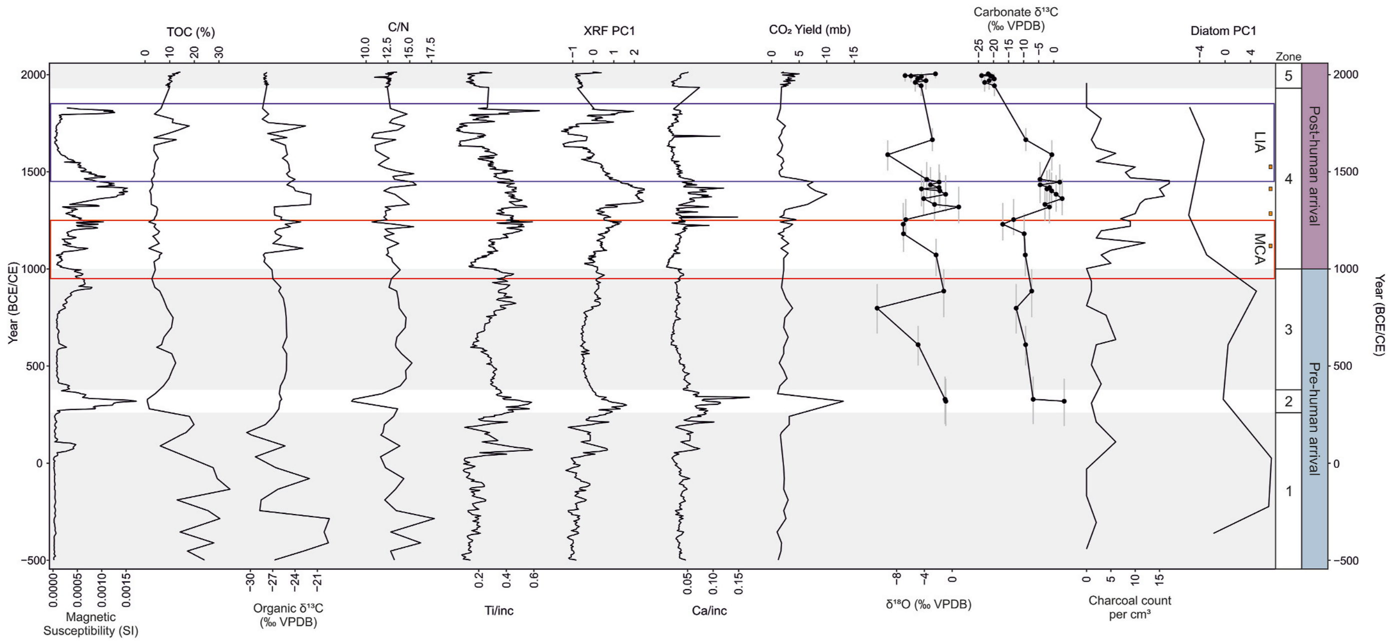

Fig. 7. Summary plot of the key proxy data from the Lake Tiriara sequence including magnetic susceptibility, organic proxies ($\delta^{13}C_{org}$), XRF data (Ti/inc, Ca/inc) and PC1 from the XRF PCA, carbonate proxies ($\delta^{18}O_{carb}$, $\delta^{13}C_{carb}$, CO_2 yield), charcoal count data and PC1 from the diatom record. The red box is showing the timing of the Medieval Climate Anomaly (MCA) and the blue box is showing the timing of the Little Ice Age (LIA) (Masson-Delmotte et al., 2014).The orange dots denote location of freshwater molluscs found within the sediments. (For interpretation of the references to colour in this figure legend, the reader is referred to the Web version of this article.)

sequence. In zone 4 the C/N values start to show a higher amplitude of change both in terms of time and C/N value.

The $\delta^{13}C_{org}$ data shows relatively high values of up to −19.4 ‰ in the lower most zone 1 but these gradually decrease and there is a large range of −30.5 to −19.4 ‰ over this zone (Fig. 7). In the latter half of zone 1, $\delta^{13}C_{org}$ start to rise and in zones 2-3 the $\delta^{13}C_{org}$ values remain relatively stable, varying between −25‰ and −27‰. From zone 4, the $\delta^{13}C_{org}$ start to fluctuate between −28.4 and −21.8‰. In a comparison of C/N against $\delta^{13}C_{org}$, the data from zone 2 show that the event layer in that zone sits within the field dominated by algal material (Fig. 8). Otherwise, the rest of the Mangaia sequence shows that the organic matter is likely derived from a mix of both aquatic and terrestrial sources.

4.4.1. Oxygen and carbon isotopes in carbonate and calcium content

The $\delta^{18}O_{carb}$ values for the Lake Tiriara sequence vary between −10.8 and + 0.9 ‰ (Fig. 7). No carbonate was present in zone 1. The $\delta^{18}O_{carb}$ in zone 2 have high values close to −1 ‰. There are two changes toward more positive values: the first between zone 3 and 4 (c. 885-1075 CE) then again later in zone 4 (c. 1320-1465 CE) beginning with the largest peak in $\delta^{18}O_{carb}$ (+0.94 ‰) around 1320 CE. In zone 5, which represents the period 1934 CE to present, the $\delta^{18}O_{carb}$ ranges from −6.78 to −2.40 ‰.

The $\delta^{13}C_{carb}$ vary between −23.97 and + 3.48 ‰. In zone 2, $\delta^{13}C_{carb}$ peaks (+3.5 ‰ at c. 320 CE) then drops and remains relatively steady through zone 3, varying between −7 and −12‰. There is a small dip in values −16.9 ‰ at c. 1230 CE before a notable step in values between 1320 and 1465 CE where values vary between −4.5 and + 2.8 ‰. The $\delta^{13}C_{carb}$ from the core top sediments (c. 1930 CE onward) are lower than those down the sediment sequence (c. −20 ‰).

The CO_2 yield data (based on the amount of CO_2 liberated during $\delta^{18}O_{carb}$ and $\delta^{13}C_{carb}$ sample preparation) shows that samples from c. 320 CE and c. 1300-1470 CE all produced high levels of CO_2, suggesting high carbonate content. Peaks in the Ca/Inc also coincide with the peaks observed in the yield data. This correlation suggests that Ca/inc is a good proxy for carbonate abundance. Overall, carbonate content is low across most of the sequence. While the isotope data are limited, the presence of carbonate in the core can tell us about the environment as the lake setting is not typically conducive to carbonate precipitation.

The $\delta^{18}O_{carb}$ and $\delta^{13}C_{carb}$ data for Lake Tiriara do not co-vary (Fig. 9). The Tiriara sediments $\delta^{18}O_{carb}$ and $\delta^{13}C_{carb}$ values provide an r^2 value of 0.18 ($p = 0.02$, $n = 27$).

4.5. Diatoms

Twelve samples from the cores were taken for diatom content (Fig. S1) to determine if there had been any major shifts in the salinity of Lake Tiriara as has been identified in the older sediments from Lake Teroto, a similar lake system to Tiriara, on the neighbouring island of Atiu (Parkes, 1994). Both lakes have an intermittent connection to the ocean through tunnels in the makatea (Ellison, 1994b; Parkes, 1994; Chagué-Goff et al., 2016). A number of samples analysed had very low total diatom counts possibly due to poor preservation. These low-count samples correspond with the major change in zone 2, the timing of human arrival at approximately 1000 CE and around 1300 CE. To pick out significant changes in the assemblage, the diatom data was put through a PCA and the PC1 score shows a major change that occurs around 1000 CE (Fig. 7). The key change in the stratigraphy is the shift from freshwater *Pseudostaurosira brevistriata* dominance prior to human arrival to a salinity-tolerant *Rhopalodia gibberula* and *Navicula phyllepta* dominance in the top half of the sequence following human settlement of Mangaia.

5. Discussion

5.1. Hydroclimate reconstruction using $\delta^{18}O$ from a small tropical lake

The oxygen isotope composition of water in lacustrine environments is primarily a function of precipitation (Leng and Marshall, 2004), modified by evaporation where ^{16}O is preferentially evaporated leading to higher $\delta^{18}O$. In tropical regions the temperatures remain relatively constant year-round and climate is defined by wet and dry seasons rather than winter and summer (Australian Bureau of Meteorology and

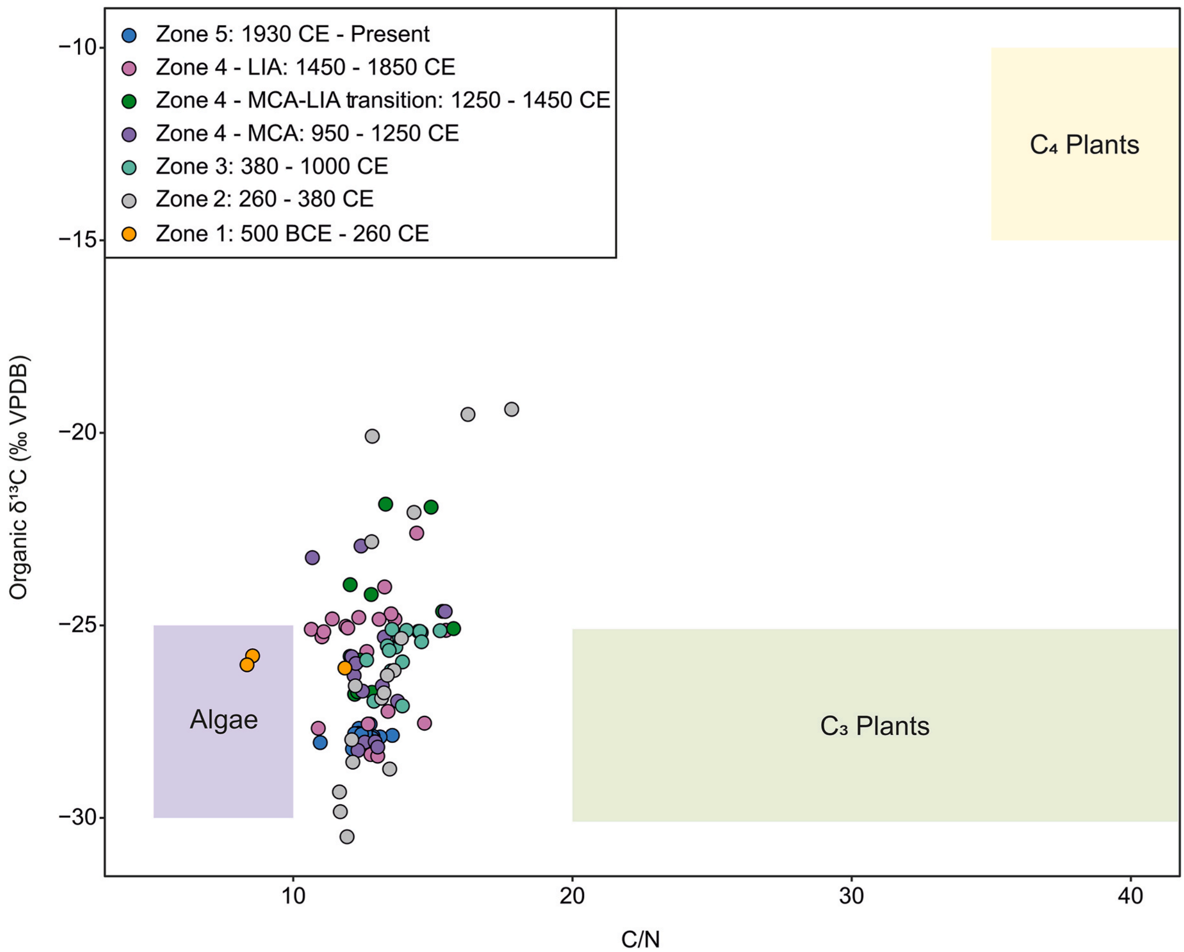

Fig. 8. Plot showing the C/N against the $\delta^{13}C_{org}$ (‰ VPDB) values for each zone of the Mangaia sequence as laid out in table SX. Boxes denote typical values for lacustrine algae (purple box), C_3 (green box) and C_4 plants (yellow box) as per Meyers (2003). (For interpretation of the references to colour in this figure legend, the reader is referred to the Web version of this article.)

CSIRO, 2011) indicating that the influence of temperature on the rate of fractionation here should be limited (Cohen, 2003; Bird et al., 2020). Carbonates are formed within the water column taking up the isotopic signature of the lake water (Lamb et al., 2005). Oxygen isotopes in carbonate are often used as a reconstruction for hydroclimate whereby higher $\delta^{18}O$ values represent drier conditions (preferential loss of ^{16}O) and lower $\delta^{18}O$ values represent wetter (fresher) conditions (Lamb et al., 2000; Bird et al., 2011; Stansell et al., 2020). Typically, hydrologically open lakes have short residence times (hours, days, weeks), a small $\delta^{18}O$ range of approximately 1-2 ‰ and so capture variations in water seasonality and temperature (Leng and Marshall, 2004). Closed lakes generally have longer residence times (months, years), with large changes in $\delta^{18}O$ of 5 to >10 ‰ suggesting the major driver is precipitation/evaporation balance (Leng and Marshall, 2004). The $\delta^{18}O_{carb}$ and $\delta^{13}C_{carb}$ (Fig. 9) from Lake Tiriara do not covary, indicating that the basin was likely not hydrologically isolated (Leng and Marshall, 2004). This could be explained by the topography of the island as Lake Tiriara sits at a low elevation and water drains radially off the high interior out towards the swampy lowlands into the lake where it drains out through the makatea. However, the sediment archive from Lake Tiriara does have a relatively large range of $\delta^{18}O_{carb}$ values (>11 ‰) for a small lake and in tropical systems such as this, the main control on $\delta^{18}O$ is evaporation (Lamb et al., 2002). Based on the range of $\delta^{18}O_{carb}$ values we suggest that Lake Tiriara has a sufficient residence time to capture evaporative processes in this sequence despite the inflow and outflow which would reduce the rate of $\delta^{18}O$ increase (Lamb et al., 2002). The lack of covariance between $\delta^{18}O_{carb}$ and $\delta^{13}C_{carb}$ is due to multiple sources of carbon (and cycling) in the lake.

In Lake Tiriara, carbonate is typically either absent or precipitated in low levels in the sediment sequence but is higher in zone 2 and 4, as conditions during these zones provided a higher concentration of ions, thereby enabling carbonate precipitation within the lake (Burn and Palmer, 2014). However, based on the range of $\delta^{18}O_{carb}$ values, whilst wetter periods are not as well preserved in the record due to the lower level of carbonate production in the lake system, there are still wetter conditions captured in the Lake Tiriara sequence where $\delta^{18}O_{carb}$ are notably lower (<−6 ‰). As such we argue that this provides confidence that the record does capture both wetter and drier periods and so provides a useful reconstruction of hydroclimate.

5.2. Overview of palaeoenvironmental change in the Cook Islands during the late Holocene

The palaeoenvironmental history of Mangaia can be split into two halves, the period prior to human arrival (550 BCE – 1000 CE) and post human arrival (1000 CE to present) (Table S2). The sequence and proxies are described in more detail in the interpretation in the ensuing sections.

5.2.1. Pre-human arrival: 500 BCE – 1000 CE

In Lake Tiriara, zone 1 (500 BCE – 260 CE) begins prior to human arrival in the Southern Cook Islands (approximately 1000 CE) (Kirch, 2017; Sear et al., 2020). At this time the site was a swampy or marshy environment as evidenced by a combination of high Br/inc with high

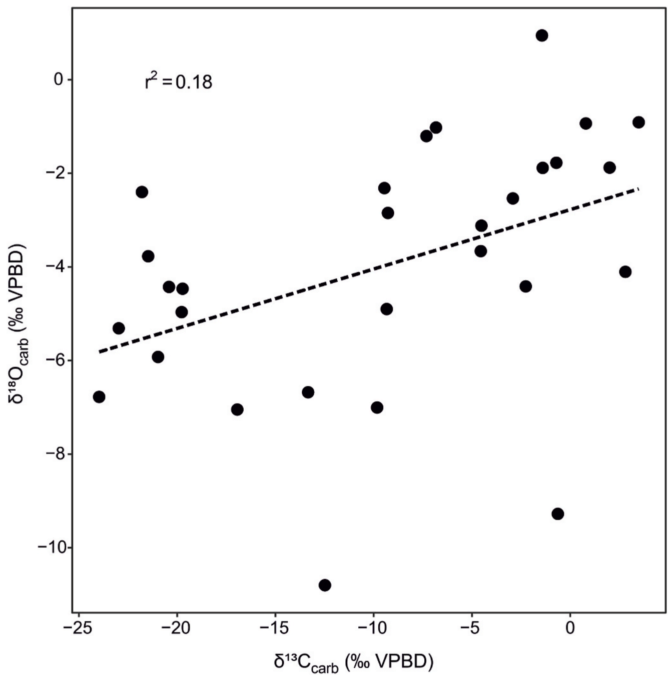

Fig. 9. Plot showing the relationship between $\delta^{13}C_{carb}$ (‰ VPBD) and $\delta^{18}O_{carb}$ (‰ VPBD) from the Lake Tiriara sediments with a regression line and the r-squared value displayed in the top left.

TOC values, which indicate more organic material/terrestrial plants (Davies et al., 2015; Cadd et al., 2018; Ribeiro Guevara et al., 2019) and the diatoms indicate standing water (Wirrmann et al., 2011). In the XRF PCA, this zone is loaded negatively on PC1 towards Br and S, both of which are used as indicators of organic matter (Moreno et al., 2007; Burn and Palmer, 2014) supporting this interpretation.

Zone 2 (260-378 CE) is unusual in comparison to the rest of the Mangaia record comprising a spike in magnetic susceptibility, Ca/inc and high $\delta^{18}O_{carb}$ and $\delta^{13}C_{carb}$ but low C/N and TOC. In the XRF PCA, this zone is loaded positively in PC1 - opposite from the zone preceding – with elements indicative of terrestrial input (Fe, Ti, Si, Rb) and carbonate input (Sr, Ca) (Chagué-Goff et al., 2016). This shows that the loading of PC1 is organic input (negative) versus terrestrial and carbonate input (positive), whilst PC2 primarily divides the terrestrial (negative) and carbonate (positive) elements. Zone 2 is the first zone that had high enough carbonate levels to produce $\delta^{18}O_{carb}$ data. There is a mismatch of signals in zone 2 as higher $\delta^{18}O_{carb}$ values and carbonate precipitation point toward drier conditions but an increase in terrestrial signals such as Ti and magnetic susceptibility indicate wetter (Sear et al., 2020; Maloney et al., 2022). One explanation is that this mix of signals could represent a tsunami event as other tsunami records show an increase in terrestrial signal (from wave backwash) (Kokociński et al., 2009; Chagué-Goff et al., 2011; Riou et al., 2020), peak in Ca/inc (Cuven et al., 2011; Chagué-Goff et al., 2016) and high $\delta^{13}C_{carb}$ (Clayton and Degens, 1959) can be indicative of a marine signal or a tsunami. Previous studies have identified potential tsunami events in the tropical South Pacific and a record from the peats surrounding Lake Tiriara suggested a potential tsunami event dated to around 150 BCE – 250 CE (Chagué-Goff et al., 2016).

In zone 3 (380 - 1000 CE), there is a steady decline in terrestrial indicators such as Ti along with a substantial drop in the Ca/inc and a rapid increase in S/inc and Br/inc. The XRF PCA (Fig. 6) indicates that the sediments in this zone are primarily negatively loaded in PC1, with Br and S, which are associated with organic content and is similarly loaded to zone 1. In other tropical lake systems, S is used as a signal for gypsum precipitation during dry periods (Burnett et al., 2011) but this is unlikely as there is not a strong relationship between S and Ca in this part of the sequence. The argument against gypsum precipitation is also supported in the XRF PCA which shows Ca and S are oppositely loaded suggesting a negative correlation between these elements (Bertrand et al., 2024). The diatom data at 370 cm from 610 CE, shows a relatively diverse assemblage where *Rhopalodia gibberula* and *Thalassiosira weissflogii* can tolerate a range of salinity conditions (Owen et al., 2004; Radchenko and Il'yash, 2006; Navarro and Lobban, 2009; Laut et al., 2019; Chen et al., 2021). However, the presence of other freshwater species such as *Pseudostaurosira brevistriata* implies a freshwater environment. This zone probably represents the evolution from a swamp to a lake system. In Eastern Australia, a similar record shows the reverse as Welby Lagoon transformed from a lake to a swamp (Cadd et al., 2018), though with the addition of brackish inwash from the makatea tunnels. Lake Tiriara is mentioned within the traditional tales from Mangaia's ethnohistory, in which it is described as a freshwater lake (Reilly, 2009) suggesting the lake has been fresh at this site within human memory.

5.2.1.1. Drought event: c. 1000 CE. There is a change in the Tiriara sequence that occurs around 1000 CE, at the onset of the Medieval Climate Anomaly (MCA) which occurred between 950 and 1250 CE (Masson-Delmotte et al., 2014). Higher $\delta^{18}O_{carb}$ values (more evaporation) indicate drier conditions from c. 885 CE (752 - 998 CE, 2σ) continuing until at least c. 1075 CE (963-1155, 2σ). This shift to a drier period aligns with the start of the MCA and the second wave of migration and human settlement into Eastern Polynesia, though there is a limited amount of $\delta^{18}O_{carb}$ data for this period. Another record from Lake Tiriara, taken from the peat at the edge of the lake, also found that around 750 CE there was a shift toward drier conditions (Chagué-Goff et al., 2016), though the limited chronological data for this record means the exact timing is uncertain. Furthermore, after human arrival, there is a notable shift in the diatom PC1 and diatom stratigraphy became dominated by *Rhopalodia gibberula* which has been identified as a species that is found in shallow lakes that experience recurrent mixing (Gasse and Fontes, 1989; Sylvestre et al., 1995; Cocquyt and De Wever, 2002; Owen et al., 2004). *Rhopalodia gibberula* has also been linked to periods of dry conditions in other records (Yacobaccio and Morales, 2005). The ecology of this species suggests it may be a useful indicator of lake shallowing and peat developing at the fringes of Lake Tiriara, suggesting it's starting to transition to a swamp as occurs in other lake systems on Mangaia (Ellison, 1994b). This corresponds with other local records, Lake Teroto on Atiu (also within the Southern Cook Islands) shows a dry phase as people were arriving into the region at c.990 CE (Sear et al., 2020). Overall, the proxies in this sequence point to a dry period just prior to or at the onset of the MCA at around 950 CE as humans are starting to move into Eastern Polynesia (Wilmshurst et al., 2008; Sear et al., 2020; Allen, 2025).

5.2.2. Post-human arrival: 1000 CE to present

According to the age model human arrival to the Southern Cook Islands is at around 306 cm (1002 CE, 875-1088 2σ) within the Lake Tiriara sequence. Zone 4 (1000-1850 CE) shows significant evidence that human activity changed the landscape. Humans had established a settlement on Mangaia by approximately 1000 CE, people had started to live permanently in the island's interior and were developing the agricultural cultigens in lowland swamp areas by 1300 CE (Kirch, 2017). Whilst most of the crops brought by Polynesians to the island are C_3 plants (Swift, 2016), the increase in the $\delta^{13}C_{org}$ could indicate a change in the type of vegetation towards C_4 plants, including grasses (Hassall, 2017), which could have colonised due to land clearance and the intensification of agriculture during this time. The theory of agricultural intensification is supported by charcoal data that shows a peak (Fig. 7) dating to c. 1450 CE (1358-1541, 2σ) along with an increase in the Ti/inc terrestrial signal. Chagué-Goff et al. (2016) suggested from their Lake Tiriara marginal peat record that an increase in terrigenous inwash (from XRF) inferred the start of wetter conditions or increased land clearance from 934 to 1111 CE. We propose that this signal is the start of a major alteration of the landscape by humans. Humans had arrived on Mangaia and established a permanent population during this time (Kirch, 2017) and the slash and burn techniques used to clear the land for agriculture by the population would increase the amount of Ti and charcoal entering the lake system. Clearing of land destabilises the surface soils and generates an increase in available material for transport when it rains (Sear et al., 2020; Maloney et al., 2022). This evidence indicates that at this time the human population had grown to such an extent that it was having a major impact on Lake Tiriara. This is most apparent in the XRF PCA, where zone 4 is positively loaded in PC1 which is associated with terrestrial input (Ti, Fe, Rb, Si) though there is a notable shift toward positive loading in PCA 1 and 2 for zone 4, separate from other zones, particularly for the transition between the MCA and the Little Ice Age (LIA) where we see the peaks in both Ti and charcoal.

The LIA sits within zone 4 and covers the period 1450-1850 CE (Masson-Delmotte et al., 2014). Other records from across the Pacific including Vanuatu, Samoa and Wallis, give mixed signals during the LIA period indicating that conditions were variable across the region, but most point to a wetter climate following the MCA-LIA transition (Maloney et al., 2022), which is supported by the Tiriara sequence. During this period, there is an increase in TOC and decrease in $\delta^{18}O_{carb}$, indicating a shift to wetter conditions though the decrease in the terrestrial inwash indicator Ti/inc would imply drier. The charcoal count drops over this period indicating a shift toward wetter conditions, i.e. less natural fire or a drop-in human related fire activity during this time. This is highlighted in the XRF PCA where the LIA sediments are less strongly weighted against the terrestrial indicators as the preceding transition zone.

Overall, there is a much lower rate of variation across all proxies in zone 5 (1934 CE - present). This highlights the importance of palaeoenvironmental studies to portray the true degree of variation occurring at sites such as Lake Tiriara. There is evidence for some remineralisation of organic carbon and an increase in CO_2 from catchment. This could be due to inwash from lake edges when they are exposed during dry periods as the lake is shallow with low gradient sloping sides (Shanahan et al., 2007; Maloney et al., 2022). Lower $\delta^{13}C_{carb}$ and $\delta^{13}C_{org}$ could be due to the rapid infilling of the lake, causing it to shallow out which leads to oxidation of the lake bottom allowing ^{12}C to escape, drawing down TDIC values (Leng and Marshall, 2004). The modern lake is less than 2 m deep and many other lakes across the island have completely been filled in (Ellison, 1994b). This interpretation is also supported by the findings of the ^{210}Pb analysis, which indicates that there has been mixing of the core top sediments altering the typical ^{210}Pb curve (Zaborska et al., 2007) (Fig. S2), this potentially could have been caused by bioturbation due to presence of freshwater eels in this shallow lake (Kirch et al., 1995) but this was likely less problematic with older sediments, when the lake would have been deeper. This zone also sees the highest Br/inc levels in the sequence, likely due to shallowing of the lake and the development of peat at its edges (Chagué-Goff et al., 2016). This is supported by the XRF PCA which shows this zone is negatively loaded in PC1 back toward the organic associated elements such as Br (Burn and Palmer, 2014).

5.2.2.1. Drought event: MCA-LIA transition c. 1300 CE. The MCA-LIA transition zone (1250-1450 CE) contains a relatively high Ca/inc and CO_2 yield indicating carbonate precipitation and as such probably represents an evaporated lake and a shift toward drier conditions (Hassall, 2017; Thompson et al., 2017). The modern lake water value of −3.3 ‰ would convert to a calculated equilibrium calcite value of −5.85 ‰ using the equation from Hays and Grossman (1991) and the reported lake water temperature of 27.6 °C from Schabetsberger et al. (2009). A majority of the $\delta^{18}O_{carb}$ data (−4.4 to +0.9 ‰) sit above the calculated modern calcite value and well above the average $\delta^{18}O_{carb}$ value for the sequence of −3.98‰ but not high enough to reflect a marine input or saline signal (Tiwari et al., 2015, Ma et al., 2017). The presence of calcium in the record due to inwash from the makatea escarpment can be ruled out based on the $\delta^{13}C_{carb}$ values, as the values for marine typically sit at around 0‰ whereas freshwater values vary more widely (Cuna et al., 2001; Oehlert and Swart, 2014) and the $\delta^{13}C_{carb}$ values from this section are around −4.6 to +2.8 ‰. An increased rate of terrestrial inwash typically would mute signals from other inputs into the system but here Ca/inc also increases, which indicates that there is a different process operating in this zone. Here the increase in the Ca/inc signal is interpreted as a drier period that reduced the lake level, concentrating the ions enough to produce a higher rate of carbonate precipitation (Conroy et al., 2008; Hassall, 2017; Thompson et al., 2017). The precipitation of carbonates such as aragonite - as shown by the strong association of Sr with Ca in the PCA - indicates that the lake has a high Mg/Ca ratio, which is driven by a high evaporation-precipitation ratio (Rosenmeier et al., 2002, Ma et al., 2017). Normally the higher rates of terrestrial inwash (Ti and magnetic susceptibility) would indicate wetter conditions (Metcalfe et al., 2010; Corella et al., 2012; Gosling et al., 2020; Sear et al., 2020), however, this does not agree with the higher $\delta^{18}O_{carb}$ values, which point to drier conditions during this period. Maloney et al. (2022) found that while the Ti and magnetic susceptibility record for Lake Lanoto'o in Samoa typically appeared to correspond to the δ^2H dinosterol changes, they became disconnected during the MCA. This is because the dry conditions that dominated during that time caused the water levels to drop resulting in an alteration to the physical characteristics of the lake morphology. The shallower margins of the lake were exposed, comprising a significant lake bench, providing an additional source of material during rainfall events. These changes resulted in a disconnection between the Ti and magnetic data erosion record and the dinosterol hydroclimate records. A study from Lake Bosumtwi in Ghana similarly found that terrestrial input into the lake was controlled by how much of the catchment was erodible. The levels of terrestrial input were higher when the water level in the lake was lower, because during drier periods, more of the lake bed is exposed, increasing the erodible catchment area (Shanahan et al., 2009). For the Lake Tiriara sequence, conditions are drier as shown by the $\delta^{18}O_{carb}$, so lake levels are lower exposing more erodible material and simultaneously humans are disturbing the catchment thus increasing Ti and magnetic susceptibility even during this dry phase. This would explain the disagreement between the $\delta^{18}O_{carb}$ and Ti/magnetic susceptibility during this transition period. A potential dry period at this time is also supported by the lack of diatoms from this zone, as only a small number of frustules were present in the sample from this zone. It has been suggested that diatom frustules can be poorly preserved in the sediments when sections of the lake dry out and the frustules are exposed to the atmosphere (Conroy et al., 2008).

In summary, the proxy data from the palaeo sequence of Lake Tiriara indicates that at the start of this sequence – zone 1 covering the period 500 BCE – 260 CE - the site was likely an organic-rich wetland. Then zone 2, covering 260-378 CE, is an event layer which we propose is a potential tsunami event preserved in this record. Zone 3 represents an evolution of the site from a wetland to a lake system. A change in the $\delta^{18}O_{carb}$ values across zones 3 and 4 suggest that there was a dry period from 885 CE until at least 1075 CE (752-1155,2σ). After the start of the human occupation of Mangaia at approximately 1000 CE (zone 4), the lake sequence is altered by human activity on the island, specifically land clearance and burning. There is evidence from the $\delta^{18}O_{carb}$ record, that there was a second dry period during the MCA-LIA transition on Mangaia between 1320 and 1460 CE (1235 - 1550 CE, 2σ). The modern part of this sequence dating from 1934 to present (1881 - 2016, 2σ) is represented in zone 5. Lake Tiriara has been rapidly infilling (Ellison, 1994b), the lake is shallowing and peat has developed at its edges. This has led to mixing within the sediments and potential oxidation of the lake bottom. The sediment archive from Lake Tiriara provides a multi-proxy palaeoenvironmental record covering approximately 2500 years and picks out key drought periods in relation to the timing of human arrival and settlement of Mangaia.

5.3. *Palaeoenvironmental change in the tropical South Pacific in the late Holocene*

Our new Mangaia palaeoclimate record corresponds with existing information from the neighbouring island of Atiu and recent regional tropical Pacific records (Toomey et al., 2016; Sear et al., 2020; Maloney et al., 2022)(Fig. 10). The Mangaia $\delta^{18}O_{carb}$ record has age model uncertainties that overlap with similar trends in the Atiu δ^2H leaf wax (Sear et al., 2020), which show that at the transition into the MCA (c. 950 CE) conditions were relatively dry and that following this, both the Mangaia $\delta^{18}O_{carb}$ and Atiu δ^2H records indicate there was a shift to wetter conditions. The controls on δ^2H in leaf waxes are not yet well understood (Gao et al., 2014) and can be affected by factors other than precipitation such as vegetation changes, biochemical isotope fractionation (Baan et al., 2023). The interpretation of hydroclimate reconstructions using leaf wax δ^2H requires an understanding of the lake catchment (Garcin et al., 2012). However, correspondence between these records which are controlled by different factors suggest that they are providing a hydroclimate signal and the record from Atiu considers changing carbon sources but there was no significant correlation between $\delta^{13}C$ and δ^2H in the n-alkanoic acids showing changing vegetation was not a controlling factor (Hassall, 2017). Corroboration from other proxies too show that the shift in diatom community from one dominated by freshwater *Pseudostaurosira brevistriata* to one dominated by more salinity-tolerant *Rhopalodia gibberula* could also indicate a possible shallowing of Tiriara or shift toward wetland conditions (Gasse and Fontes, 1989; Sylvestre et al., 1995; Cocquyt and De Wever, 2002; Owen et al., 2004; Yacobaccio and Morales, 2005) due to dry conditions occurring around human arrival to the island. The Toomey et al. (2016) record (Fig. 10-C) shows more subtle variations. There is an overall upward shift in the average value of the Ti/Ca at around 950 CE, but the Mangaia and Atiu records both indicate wetter conditions within the MCA following the initial dry conditions. However, Ti/Ca is not a direct precipitation proxy, as it can be influenced by erosion and marine processes. This could represent an increase in human activity on the island. Discounting the Toomey record, the other hydroclimate records from the eastern Polynesian region indicate that there was likely a shift to drier conditions in the lead up to the second wave of human migration across the Pacific into Eastern Polynesia. Furthermore, several records from Melanesia and western Polynesia also indicate a shift to drier conditions during this time indicating that this was a climatic shift that impacted large areas of the tropical South Pacific. For example, the Samoan $\delta^2H_{dinosterol}$ record

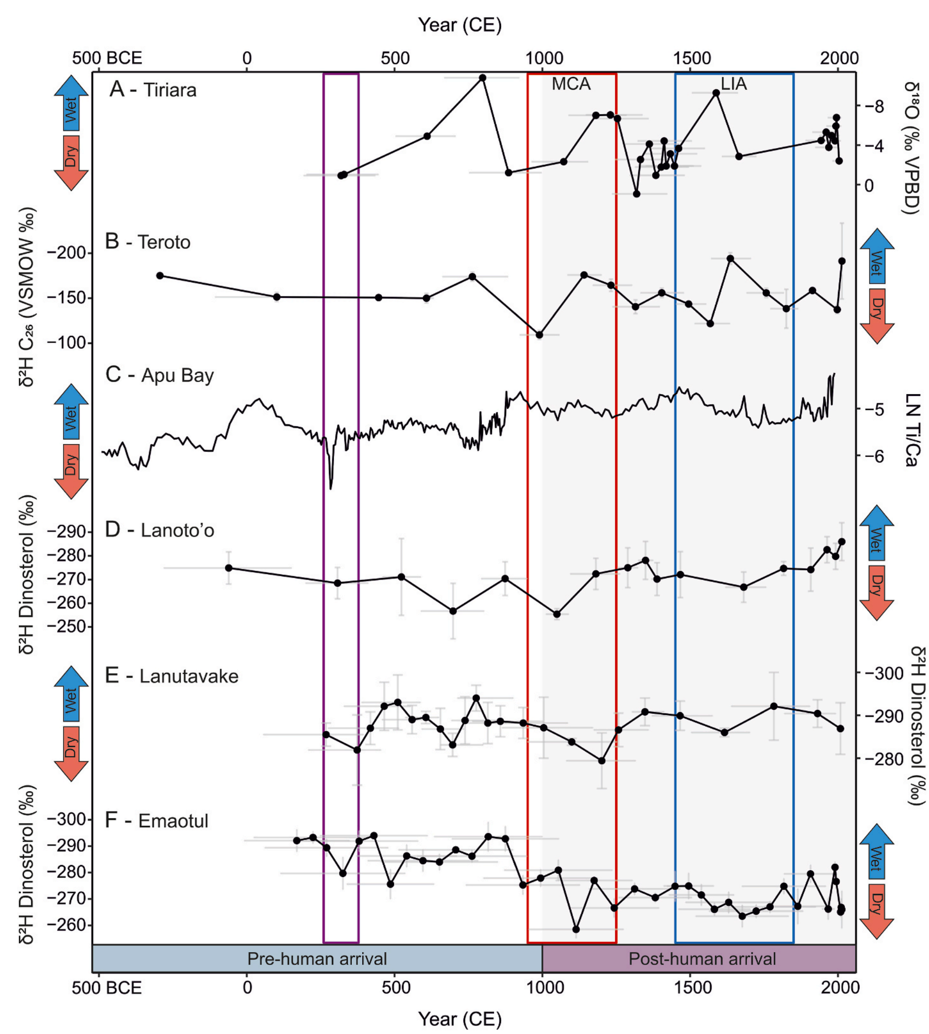

Fig. 10. Plot showing a comparison between palaeo records from the tropical South Pacific relating to hydroclimate change over the past 2500 years. A - Lake Tiriara $\delta^{18}O_{carb}$ record from Mangaia, Southern Cook Islands (this study), B – δ^2H C_{26} alkanoic acid (‰ VSMOW) record from Lake Teroto - Atiu, Southern Cook Islands (Sear et al., 2020), C – Ti/Ca record from Apu Bay, Society Islands (Toomey et al., 2016), D – δ^2H Dinosterol (‰ VSMOW) record from Lake Lanoto'o – Upolu, Samoa (Sear et al., 2020) E – δ^2H Dinosterol (‰ VSMOW) record from Lake Lanutavake – Uvea, Wallis and Futuna (Maloney et al., 2022) and F – δ^2H Dinosterol (‰ VSMOW) record from Lake Emaotul – Efate, Vanuatu (Sear et al., 2020; Maloney et al., 2022).The red and blue rectangles denote the MCA and LIA periods respectively. The purple rectangle denotes a possible event layer from the Mangaia sequence. Arrows on the plots indicate the direction associated with shifts between drier and wetter conditions from each of the data sets. (For interpretation of the references to colour in this figure legend, the reader is referred to the Web version of this article.)

from Sear et al. (2020) (Fig. 10-D), the Lake Lanutavake $\delta^2H_{dinosterol}$ record (Fig. 10-E) from Maloney et al. (2022) and a $\delta^2H_{dinosterol}$ record from Efate on Vanuatu (Sear et al., 2020; Maloney et al., 2022) (Fig. 10-F), all show a period of drying leading up to and surrounding the second wave of migration. There has been recent growth in the use of $\delta^2H_{dinosterol}$ as a hydroclimate proxy in the region. The advantages of this proxy is that it is able to provide quantitative rainfall reconstructions, which are based on a South-Pacific based calibration data set (Maloney et al., 2019). Whilst the errors can be large and the δ^2H from dinosterol can be affected by environmental conditions (salinity, temperature, light) and species (composition, growth rate, metabolism) (Maloney et al., 2019) though these issue are minimised through careful site selection and the utilisation of archives from across the South Pacific region (Maloney et al., 2019, 2022). The agreement of the $\delta^{18}O_{carb}$ record from Lake Tiriara with other records within the same locality but also within the wider region provides confidence in the interpretation of the $\delta^{18}O_{carb}$ as well as $\delta^2H_{dinosterol}$ as a hydroclimate signal. The palaeo records from Tiriara and across the South Pacific demonstrate that a regional drying phase occurred around 1000 CE, coinciding with the second wave of migration into Eastern Polynesia, after which there was a shift back towards wetter conditions towards the end of the MCA.

One of the key findings from the Mangaia sequence was the occurrence of a second dry period in zone 4 occurring between 1320 and 1460 CE (1215-1565 CE, 2σ). There is evidence for this second dry period about 300-400 years after the initial settlement of Eastern Polynesia across multiple records. The $\delta^{18}O_{carb}$ record from Lake Tiriara and the δ^2H C_{26} leaf wax from Atiu (Sear et al., 2020), show a notable shift toward dry conditions in the transition period between the MCA and the LIA (Fig. 10-B). This shift to dry conditions is also seen in other records in the Pacific such as Vanuatu (Maloney et al., 2022), Samoa (Hassall, 2017; Maloney et al., 2022) though the timing of this shift appears to be later during the LIA. Maloney et al. (2022) found there was an almost 20 ‰ drop in dinosterol values in a lake sequence from Efate in Vanuatu, equivalent to a 2 mm/day change starting at the MCA-LIA transition and across the LIA (Fig. 10-F). Higgins et al. (2020) found that the period 1310 to 1330 CE saw the largest eastward shift in the SPCZ, which was larger than the major dry period observed at the start of the MCA. However, the records used by Higgins et al. (2020) were all from marginal sites around the Pacific basin rather than the SPCZ zone, calling into question the applicability in this context. Nevertheless, this evidence all shows that around 1300 CE during the transition between the MCA and the LIA there was a shift to drier conditions in the tropical Pacific that may have persisted into the start of the LIA, though the precise timing of these changes needs refining.

There is a spatial divergence between the regional records across the LIA. The Mangaia and Atiu records show a similar pattern of rainfall change with the LIA starting with relatively drier conditions before shifting towards wetter conditions in the latter half of the period. Whereas the Samoan record shows a continuous drying trend throughout the LIA, only from 1816 CE onwards do conditions start to get wetter, which persists through to the present day. However, the error bars associated with the data from Samoa mean that there is a higher level of uncertainty around the direction of change (Maloney et al., 2022) as the range of possible values could potentially be lower indicating a drop rather than an increase in the precipitation rates. Conversely, the errors on the $\delta^{18}O_{carb}$ and dinosterol record from Atiu are much smaller, meaning greater relative confidence can be ascribed to these records.

One weakness of these palaeo archives in the South Pacific region is that all the age models from the records presented in Fig. 10 have centennial-level errors associated with them. This means that the exact onset, duration and scale of hydroclimate change within the South Pacific region remains uncertain. This highlights the need for new high-resolution records with robust chronological models to refine the timing of climatic change in this region. Despite this, there is still some agreement on the timing of major changes particularly around 1000 CE, so these records provide valuable insight to hydroclimate change in the region, but this still requires further work to refine chronologies.

In summary, over the course of the late Holocene, taken together and accounting for analytical and age model uncertainties, the combined regional records point to a shift to dry conditions around the time of the second wave of migration into the east Pacific. There is also evidence for a second dry period that started in the transition period between the MCA and the LIA and persisted into the beginning of the LIA.

5.4. Holocene climate change in the Southern Cook Islands, human migration and human settlement of Mangaia

Some of the key climate and environmental indicators from the Mangaia palaeo record are shown in Fig. 11 alongside archaeological findings in relation to the key archaeological phases of the Mangaian record as laid out by Kirch (2017). Human arrival on Mangaia at the start of the Vairorongo phase (1000-1300 CE) likely occurred within the first dry period identified in the Lake Tiriara sequence at approximately 885 - 1075 CE. Kirch (2017) suggested that this period could have been a phase of intermittent human occupation. Mangaia may have been a source of additional resources for populations of neighbouring islands during this dry period. There is an indication in the archaeological record of a disturbance occurring around 1000 CE that was felt across multiple island groups in Melanesia and Western Polynesia. In the lead up to this period there is a shift towards monument building and the inland expansion of settlements and population growth in Samoa (Quintus and Cochrane, 2018; Harris et al., 2020), a period of social contraction in Fiji (Cochrane, 2018a), monument and fortification building in Tonga (Kirch, 1984; Burley and Addison, 2018) and the rise of the Traditional Kanak cultural complex in New Caledonia (Sand, 2018). Considering this archaeological evidence of societal change alongside the evidence from multiple palaeoenvironmental archives including the Tiriara record discussed above, there is support for the hypothesis that there was a significant shift in the hydroclimate that could have had major implications for island populations around 1000 CE in the Western South Pacific that potentially led to migration into Eastern Polynesia.

A permanent population was in place on Mangaia by the start of the Tangatatu Phase (1300-1400 CE). As discussed, the palaeo record from Lake Tiriara identifies a second dry period occurring around 1320 - 1460 CE This dry period corresponds with evidence for an increase in conflict in Mangaia between 1390 CE and 1470 CE (Fig. 11). Kirch (2017) suggested that this indicated that there was pressure on the population during this period, possibly due to a resource depression that occurred between the Tangatatau (1300-1400 CE) and Ngaaitutaki (1400-1600 CE) phases in the Tangatatau Rockshelter sequence. Recent modelling work also shows how the timing of this dry period would have likely have hit as the population was reaching the islands carrying capacity leading to a severe impact on the population due to low food availability (Hipkiss et al., 2025). The sustained period of Ti input into the lake following human arrival, which indicates a higher extended rate of erosion from the catchment likely due to clearance of the inner volcanic slopes leading to catchment disturbance (Fig. 11). This is also supported by the peak in charcoal at the boundary between the Tangatatau and Ngaaitutaki phases that is used as an indicator of land clearance by humans (Gosling et al., 2020; Strandberg et al., 2023). This fits with the archaeological record as Kirch (2017) suggests that the population had peaked by the start of the Ngaaitutaki phase. The end of the Tangatatau phase also sees a peak in the rats utilised within the Mangian diet perhaps as other resources became limited. The new palaeo data alongside the archaeological record shows that this period sees the culmination of compounding pressures on Mangaia - the driest conditions seen within the time of human occupation, the height of land clearance with associated erosion and the likely peak in island population. Collectively, these pressures may have led to resource stress resulting in the outbreaks of conflict known to have occurred on

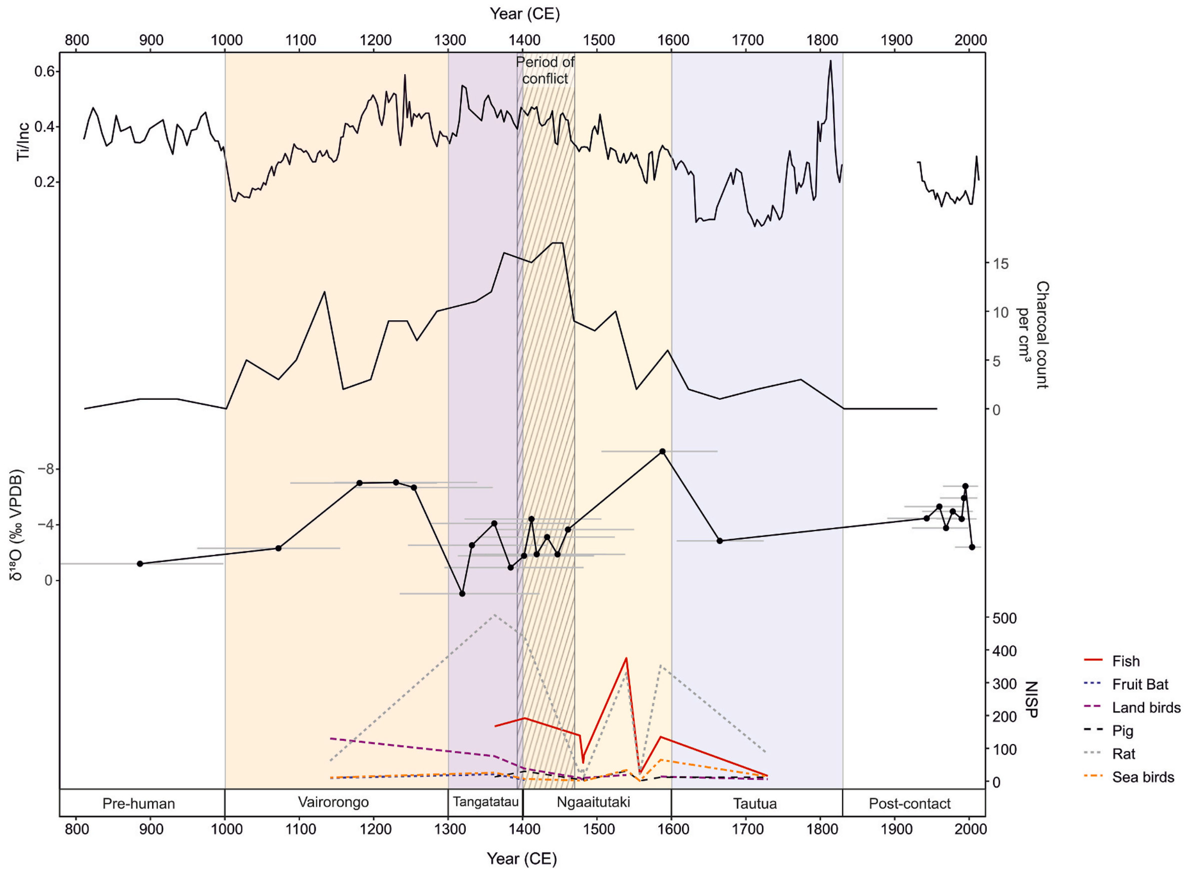

Fig. 11. Plot comparing the Ti/inc erosion data, charcoal counts and $\delta^{18}O_{carb}$ hydroclimate reconstruction data from the Lake Tiriara sediments shown against archaeological periods prescribed by Kirch (2017), faunal number of identified specimens (NISP) data from the Tangatatau rockshelter and the shaded block identifies periods of conflict as identified by (Kirch, 2017).

Mangaia (Kirch, 2017). The reality is that by approximately 1300 CE humans had rapidly expanded out and settled on the uninhabited islands of the Pacific (Wilmshurst et al., 2011; Ioannidis et al., 2021) so there was no further mechanism to relieve stress from over-population by migration. Previously, islanders had utilised resources from their own islands as well as smaller neighbouring islands and exchanged resources through inter-island networks (Allen, 2025). However, between 1300 and 1600 CE there was a notable drop in long-distance voyaging between islands (Weisler et al., 2016). When all islands are populated and reaching the absolute island capacity and there is nowhere to go, islanders have turned to other mechanisms in order to release some of that pressure of rapid growth. Other societal mechanisms to cope with resource pressure are also seen across Pacific islands around 1300 – 1600 CE including building fortifications to defend key resources (Marais, 1990; Best, 1993; Field, 2004; Field and Lape, 2010) and outbreaks of conflict (Nunn, 2000; Field and Lape, 2010; Goff and Nunn, 2016). Changes in climate and rapid population growth with no mechanism for relief may have altered human behaviour and their priorities to focus on developing agriculture and protection of resources (Allen and Craig, 2009).

By the start of the final Tautua phase (1600-1830 CE) Mangaia had gone through a period of agricultural intensification and had transformed the lowland swamp areas into productive taro pondfields (Kirch, 2017). The palaeo record shows that just before the start of the Tautua phase there was an increase in rainfall that may have provided some relief and assistance with the agricultural intensification, but it was potentially short-lived as $\delta^{18}O_{carb}$ values dip again in the Tautua phase. This phase also sees the end of the higher Ti and charcoal levels, which aligns with Kirch's (2017) suggestion that most of the landscape in the fertile lowlands had been transformed into taro pondfields by this point, capturing the inwash from the higher in-land slopes. The comparison of the palaeo and archaeological records shows how the archaeological record of human arrival and settlement can be linked to the climatic and environmental changes identified in palaeo archives providing useful context to Polynesian prehistory.

6. Conclusions

Here we provide a new multi-proxy palaeo-environment record from the tropical South Pacific covering approximately the past 2500 years including one of the first carbonate $\delta^{18}O$ records from lake sediments in the region. This provides the opportunity for comparison to the local archaeology to identify climate-environment-human interactions over the past 1000 years following human arrival into Eastern Polynesia. The key findings from the Mangaia sequence are the presence of two shifts towards dry conditions from around 885 CE until at least 1075 CE, which coincides with the arrival of humans to Mangaia and from around 1320 CE until 1460 CE. A comparison between different hydroclimate records in the region have allowed for a wider assessment of climatic change during this period finding similarities across multiple records surrounding the second wave of human migration but a lack of agreement between records in the period following the MCA. We propose that this was a regional change in the tropical South Pacific likely related to changes in the location or extent of the SPCZ and the associated shifting patterns of precipitation.

Author contributions

Our article “Connecting climate and human settlement in the tropical South Pacific: Insights from a lake sediment archive from the Southern

Cook Islands" is original work by the authors and has not been published elsewhere. All authors have made substantial contributions to submission and have approved the final version of the manuscript. Author contributions were as follows: Charlotte V. Hipkiss: Conceptualization, Investigation/Data collection, Formal analysis, Visualization, Writing - Original Draft. Justin Sheffield: Conceptualization, Writing - Review & Editing, Supervision. David Sear: Conceptualization, Writing - Review & Editing, Supervision. Peter Langdon: Conceptualization, Writing - Review & Editing, Supervision Melanie Leng: Methodology, Formal analysis, Writing - Review & Editing. Constance Temple-Brown: Investigation/Data collection.

Declaration of competing interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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Appendix A. Supplementary data

Supplementary data to this article can be found online at https://doi.org/10.1016/j.quascirev.2026.110006.

Data availability

The data presented in this paper is available from the University of Southampton data repository listed under Tiriara_Data_REPOSITORY - https://doi.org/10.5258/SOTON/D2879.

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