Reduced oxygenation at intermediate depths of the southwest Pacific during the last glacial maximum

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Abstract

To investigate changes in oxygenation at intermediate depths in the southwest Pacific between the Last Glacial Maximum (LGM) and the Holocene, redox sensitive elements uranium and rhenium were measured in 12 sediment cores located on the Campbell and Challenger plateaux offshore from New Zealand. The core sites are currently bathed by Subantarctic Mode Water (SAMW), Antarctic Intermediate Water (AAIW) and Upper Circumpolar Deep Water (UCDW). The sedimentary distributions of authigenic uranium and rhenium reveal reduced oxygen content at intermediate depths (800-1500 m) during the LGM compared to the Holocene. In contrast, data from deeper waters (≥ 1500 m) indicate higher oxygen content during the LGM compared to the Holocene. These data, together with variations in benthic foraminiferal $\delta^{13}\text{C}$, are consistent with a shallower AAIW-UCDW boundary over the Campbell Plateau during the LGM. Whilst AAIW continued to bathe the intermediate depths (≤ 1500 m) of the Challenger Plateau during the LGM, the data suggest that the AAIW at these core sites contained less oxygen compared to the Holocene. These results are at odds with the general notion that AAIW

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was better oxygenated and expanded deeper during the LGM due to stronger westerlies and colder temperatures. These findings may be explained by an important change in AAIW formation and circulation.

1. Introduction

The oxygen content in the ocean interior is determined by the balance between the supply of oxygen by ventilation (the process whereby surface mixed layer water is transported into the ocean interior) on one hand and its removal by bacterial respiration of labile organic matter, associated with remineralisation, on the other. Observations have shown that the oceans have lost 2% of their total oxygen content since 1960 (Schmidtke et al., 2017) and that the rate of deoxygenation has been increasing (Helm et al., 2011). Potential collapse of fisheries and enhancement of global warming in response to this worldwide increasing marine deoxygenation pushed the scientific community to focus on understanding the marine oxygen cycle (Helm et al., 2011; Diaz and Rosenberg, 2008; Keeling et al., 2010; Matear et al., 2000; Bograd et al., 2008; Nevison et al., 1995). Despite this greater attention over the last decade, the complex consequences of climatic forcing on oceanic oxygenation prevent future oxygenation projections from being made accurately (Emerson and Bushinsky, 2014).

This study focuses on the southwest Pacific sector of the Southern Ocean. Today, oceanic productivity in this region is relatively low (Murphy et al., 2001), thus ventilation and 'upstream' oxygen removal related to organic matter remineralisation constitute the primary factors controlling dissolved oxygen levels. Ventilation in this region occurs mainly through the formation of Subantarctic Mode Waters (SAMW) and Antarctic Intermediate Waters (AAIW), two oxygen rich water masses present at intermediate depths in the Southern Ocean (600-1200 m and 500-600 m, respectively; Figure 1, 2). They are formed by subduction of Antarctic surface waters below buoyant subtropical waters (Talley, 2013). Prior to their subduction, their pre-formation is controlled by the Ekman driven upwelling of oxygen-depleted Upper Circumpolar Deep Water (UCDW, 1500-2500 m) and Lower Circumpolar Deep Water (LCDW, 2500-3000 m) under the influence of westerly winds (Figure 2) (Sloyan and Rintoul, 2001).

Upper Circumpolar Deep Water and LCDW become oxygenated through atmospheric exchanges when they reach the surface. Then, they are advected and subducted northward, feeding the AAIW and SAMW (Talley, 2013). Of the two water masses, AAIW is volumetrically the largest and dominates the oxygen supply to the ocean interior at low latitudes (Piola and Georgi, 1982). Bostock et al. (2013) showed that there are three types of AAIW present around New Zealand (Figure 1). There is a southern source AAIW, formed locally in the southwest Pacific sector of the Southern Ocean as well as an AAIW coming from the southeast Pacific via the South Pacific gyre. Finally, the third type, which is present in the Tasman Sea, is a mixture between the AAIW from the southeast Pacific source and surface waters from the Tasman Sea. Because AAIW has an essential role in ventilating the ocean interior, modelling studies have focused on AAIW to predict future oxygen variations (Sallee et al., 2010; Rintoul and Bullister, 1999; Downes et al., 2009). However, large uncertainties remain about the mechanisms driving AAIW formation in the modern ocean (Bostock et al., 2013). Therefore, more work needs to be done in order to understand how our changing climate will influence AAIW formation and its oxygenation role.

A precise historical knowledge of how AAIW responded to past climatic forcing is essential to help understand how AAIW formation will change in the future. Hence, the reconstruction of past variations in oxygenation may be useful to better predict future oxygen changes in the Southern Ocean. Previous studies provided conflicting results regarding the AAIW geometry in the Pacific sector of the Southern Ocean between glacial and interglacial periods. Using sponge $\delta^{30}\text{Si}$ (Rousseau et al., 2016) or a combination of benthic foraminiferal $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ (Elmore et al., 2015; Pahnke and Zahn, 2005; Ronge et al., 2015), several authors reported AAIW contraction during glacial periods in the southwest Pacific sector of the Southern Ocean. They proposed that this AAIW contraction decreased the ventilation at intermediate depths. However, other authors reported increased ventilation at intermediate depths in the southeast Pacific sector of the Southern Ocean (along the Chilean margin) during the last ice age (Muratli et al., 2009). They attributed this increased ventilation to greater AAIW formation. Based on a large scale data compilation, Jaccard and Galbraith (2012) and Jaccard et al. (2014) showed that intermediate depths of the Pacific Ocean were

Figure 1: Map of the bathymetry of the New Zealand region as well as the sediment core locations on the left side. The present AAIW circulation is also shown (Chiswell et al., 2015). Three different AAIW types are present in the New Zealand region: AAIW 1 is a southern source AAIW, formed locally in the southwest Pacific sector of the Southern Ocean. AAIW 2 comes from the southeast Pacific via the South Pacific gyre. AAIW 3 is a mixture between the AAIW from the southeast Pacific source and surface waters from the Tasman Sea. The position of the P15S section used in Figure 2 is also represented. On the right side is shown the bathymetric transect from 180°W.

generally better oxygenated during the Last Glacial Maximum (LGM) compared to the Holocene, while abyssal waters remained poorly ventilated. In the rest of the Southern Ocean, several studies also showed a poorer ventilation during the LGM compared to the Holocene (Wagner and Hendy, 2017; Lu et al., 2016; Frank et al., 2000; Jaccard et al., 2016); however, a general lack of data limits interpretations for this region. Consequently, uncertainties remain about past oxygen variations in the Pacific sector of the Southern Ocean. In particular, redox-sensitive metal proxies of bottom water oxygen have not been applied yet in the southwest Pacific.

In this study we aim to reconstruct the variations in the oxygen content of the intermediate waters of the southwest Pacific sector of the Southern Ocean, between the LGM and the Holocene. This region is of particular interest because parts of the shallow plateaux surrounding New Zealand are bathed by AAIW and are therefore ideal

Figure 2: Oceanographic settings of the New Zealand region. Dissolved oxygen concentrations from the P15S (a) and P11 (b) sections (WOCE). The positions of the sediment cores analysed (aU, aRe) on the Campbell and Challenger plateaux are represented on the top and bottom panels respectively. (c) Carbon isotope composition of Dissolved Inorganic Carbon ($\delta^{13}C_{DIC}$) from the P15S section of the WOCE. The positions of the sediment cores from which benthic foraminifera $\delta^{13}C$ were used in this study were added, although some of these cores are from the Tasman Sea, the $\delta^{13}C$ is assumed to be similar due to the relatively similar oxygen profiles in A and B. A simplified Southern Ocean circulation is shown from Talley (2013). The northern and southern AAIW sources are represented from Bostock et al. (2013). Graphs were plotted using Ocean Data View (Schlitzer, 2017)

70 to investigate AAIW changes (Figure 1 and 2) (Bostock et al., 2013; Forcén-Vázquez et al., 2017; Chiswell et al., 2015).

To this end, the authigenic Uranium (aU) and Rhenium (aRe) contents of 12 sediment cores from the New Zealand region were analysed (Campbell Plateau, Challenger Plateau, Bounty Trough) (Figure 1, 2, Table 1). The solubility of U and Re in seawater is dependent on the seawater oxygen concentration. When the dissolved oxygen concentration decreases, U and Re are reduced from their soluble forms ($UO_2(CO_3)_3^{4-}$, $U(OH)_4$ and ReO_4^-) to insoluble forms ($UO_{2(s)}$ and $ReO_2 \cdot 2H_2O$). This causes aU and aRe solid phases to accumulate in the sediments (Colodner et al., 1993; Morford and Emerson, 1999; Tribovillard et al., 2006; Crusius et al., 1996). This relationship
80 between aU and aRe sediment enrichments and oxygen concentration is not linear and displays more of a threshold response (Zheng et al., 2002a,b; McManus et al., 2005, 2006). Consequently, aU and aRe concentration variations in sediment are a semi-quantitative proxy for oxygenation changes. In the Southern Ocean, previous studies have reported that low concentrations of aU and aRe in sediments, less than 1 mg/kg and
85 5 $\mu\text{g/kg}$ respectively, correspond to well oxygenated waters, while high concentrations of aU and aRe, around 4 mg/kg and 20 $\mu\text{g/kg}$ respectively, correspond to water with low oxygen content, without necessarily reaching anoxia (Chase et al., 2001; Muratli et al., 2009; Chase et al., 2003; Hayes et al., 2014; Gottschalk et al., 2016; Jaccard et al., 2016). In order to identify potential drivers of oxygen variations, circulation
90 changes were also reconstructed using available stable carbon isotopes from benthic foraminifera ($\delta^{13}\text{C}$ benthic) from 8 sediment cores in the studied region (Figure 1, 2, Table 1).

Table 1: Summary of the sediment core locations, depths, proxies used and age models. Delta ^{13}C data all correspond to benthic foraminifera $\delta^{13}\text{C}$. An indication of the water masses, bathing the core sites is included, based on the densities of the P15S section from the WOCE. When the core is at the interface of two water masses, both are reported. (*) Previously published age models were used without modifications (***) New radiocarbon dates were used to update previously published age models.

Core	Latitude	Longitude	Depth (m)	Location	Present water Mass bathing core site	Proxy used	References	Age model
DSDP 593	-40.5	167.67	1068	Challenger P.	AAIW	aU, aRe, $\delta^{13}\text{C}$	aU and aRe: This study; $\delta^{13}\text{C}$: Elmore et al. (2015)	Radiocarbon (this study) + $\delta^{18}\text{O}$ alignment with LR04 stack (Martinez et al. 1994)
MD06-2986	-43.45	167.9	1477	Challenger P.	UCDW	$\delta^{13}\text{C}$	Ronge et al. (2015)	(*) Ronge et al. (2015)
MD06-2988	-42.35	167.93	1792	Challenger P.	UCDW	aU, aRe	This study	(*) Neil, NIWA (unpublished)
MD06-2990	-42.32	169.92	943	Challenger P.	AAIW	$\delta^{13}\text{C}$	Ronge et al. (2015)	(*) Ronge et al. (2015)
MD97-2120	-45.53	174.93	1210	Campbell P.	AAIW	$\delta^{13}\text{C}$	Palmke et al. (2005)	(*) Palmke et al. (2005)
Q217	-46	175	1936	Campbell P.	UCDW	aU, aRe	This study	Radiocarbon (this study)
SO136 GC38	-50.13	175.18	1359	Campbell P.	AAIW-UCDW	aU, aRe	This study	(*) Neil et al. (2004)
SO136-003	-42.3	169.88	958	Challenger P.	AAIW	$\delta^{13}\text{C}$	Sturm (2003)	(*) Sturm (2003)
SO213-82-1	-45.77	176.6	2066	Campbell P.	UCDW	$\delta^{13}\text{C}$	Ronge et al. (2015)	(*) Ronge et al. (2015)
SO213-84-1	-45.12	174.57	972	Campbell P.	AAIW	$\delta^{13}\text{C}$	Ronge et al. (2015)	(*) Ronge et al. (2015)
TAN0803-09	-47	166.06	1648	Campbell P.	UCDW	aU, aRe	This study	(*) Bostock et al. (2015)
TAN1106-11	-47.17	166.65	1349	Campbell P.	AAIW-UCDW	aU, aRe	This study	(Bostock, NIWA; unpublished) Radiocarbon (this study) + $\delta^{18}\text{O}$ alignment with LR04 stack
TAN1106-7	-46.77	166.92	743	Campbell P.	AAIW	aU, aRe	This study	(Bostock, NIWA; unpublished) Radiocarbon (this study) + $\delta^{18}\text{O}$ alignment with LR04 stack
Y1439	-52.91	168.12	791	Campbell P.	AAIW	aU, aRe, $\delta^{13}\text{C}$	aU and aRe: This study; $\delta^{13}\text{C}$: Neil, NIWA (unpublished)	Radiocarbon (Neil et al. 2004) + $\delta^{18}\text{O}$ alignment with LR04 stack (Neil et al. 2004)
Y16	-50.59	169.76	600	Campbell P.	AAIW	aU, aRe	This study	Radiocarbon (***) + $\delta^{18}\text{O}$ alignment with LR04 stack (Neil et al. 2004)
Y17	-48.39	169	691	Campbell P.	AAIW	aU, aRe	This study	Radiocarbon (***) + $\delta^{18}\text{O}$ alignment with LR04 stack (Neil et al. 2004)
Y8	-46.58	178.39	1335	Campbell P.	AAIW-UCDW	aU, aRe	This study	(*) Ellwood et al. (2005)
Y9	-48.24	177.34	1267	Campbell P.	AAIW-UCDW	aU, aRe	aU and aRe: This study; $\delta^{13}\text{C}$: Neil, NIWA (unpublished)	Radiocarbon (Neil et al. 2004) + $\delta^{18}\text{O}$ alignment with LR04 stack (Neil et al. 2004)

2. Methods

2.1. Age models

95 Previously published age models were used without modifications for several sediment cores in this study (Table 1). For the other cores, mixed planktonic foraminiferal radiocarbon data were used to update or to construct new age models. The new radiocarbon measurements were conducted by the ANSTO Science Institute (Lucas Heights, Australia) using an accelerator mass spectrometer (AMS). All AMS dates (new and
100 pre-existing) were calibrated with the MARINE13 calibration curve (Reimer et al., 2013) using the OxCal 4.2 platform (Ramsey, 2009) and a modelled local reservoir age of 500 years (Skinner et al., 2017). When foraminiferal $\delta^{18}O$ were available (Table 1), the radiocarbon and foraminiferal $\delta^{18}O$ were combined to construct/update the age models. The radiocarbon dates were used as tie points, and correlated the $\delta^{18}O$ records
105 to the LR04 stack using the Analyseries software (Paillard et al., 1996). Radiocarbon ages and uncertainties used to develop new age models are presented in Table 2.

2.2. Sediment preparation and digestion

Sixty mg of freeze-dried and homogenized sediments were digested in a mix of HCl, HNO_3 and HF using a pressure-assisted microwave (Milestone Ethos SK-12
110 microwave oven, Milestone, Shelton, CT, USA). The sample preparation has been described in detail previously in Durand et al. (2016).

2.3. Thorium, Uranium and Rhenium Analysis

The day prior to analysis, samples were heated on a hot plate at 60 °C for 24 hours and then diluted by a factor of 10 with milli-Q water. At this stage Indium (^{115}In) was added
115 as an internal standard. An ELEMENT 2 Sector Field Inductively Coupled Plasma-Mass spectrometer (ICP-MS) from Thermo Fischer Scientific (Bremen, Germany) was used in low resolution ($m/\Delta m \approx 400$) to measure Re, Thorium (Th) and U concentrations, with isotopes ^{115}In , ^{185}Re , ^{232}Th and ^{238}U monitored. An external calibration method was used for the elemental quantification. The calibration solutions were prepared daily
120 from 100 $\mu g/mL$ standard solutions (QCD Analysts, Spring Lake, NJ, USA). Further

Table 2: Radiocarbon ages and uncertainties used to develop new age models.

Cores	Uncalibrated ages BP (years)	Calibrated age BP (years)
DSDP 593	6940±70 at 0.5 cm	7364±274
	10430±45 at 15.5 cm	11881.5±427
	16980±70 at 27.5 cm	20033±451
	19750±90 at 45.5 cm	23346±538
Q217	9430±45 at 24.5 cm	10213±254
TAN1106-11	1570±60 at 1.5 cm	1999±259
	11030±70 at 61.5 cm	12462±310
	13690±60 at 121.5 cm	16103±519
TAN1106-7	5455±35 at 1.5 cm	5816±110
	10070±60 at 41 cm	11209±631
	16260±80 at 111 cm	19228±510
Y16	4280±40 at 0.5 cm	4405±255
	7400±40 at 24.5 cm	7773±274
	36320±290 at 48.5 cm	40504±1254
Y17	3460±30 at 0.5 cm	3302±189
	7755±40 at 36.5 cm	8082±164
	18360±80 at 56.5 cm	21762±472

details concerning sample handling procedures, as well as the instrument operating conditions can be found in Durand et al. (2016). The long-term reproducibility of the method was tested over multiple digestions and analyses of D-178, a surface sediment from the Campbell Plateau. Across the 19 measurements over a three month period,
125 standard deviations observed for Re, Th and U were less than 10% (Durand et al., 2016).

2.4. Calculation of authigenic Uranium and Rhenium concentrations

The total U and Re concentrations in the sediments represent input from both the land (detrital U and Re) and the water (authigenic U and Re). Authigenic U, and Re (aU and aRe) concentrations were estimated as follows (M indicates Metal):

$$M_{authigenic} = M - \left(\frac{M}{Th}\right)_{cc} \cdot Th_{total} \quad (1)$$

130 The mean continental crust (cc) ratios $(U/Th)_{cc}$ and $(Re/Th)_{cc}$ used were 0.23 and $33e^{-6}$ respectively (Rudnick and Gao, 2003). Mean continental crust ratios were compared to the records available in the New Zealand region (EarthChem database) and the averaged crust ratios were found to be representative for the New Zealand region. Overall, the detrital fraction represented approximately 10 to 25% of the total U for
135 the cores located on the Campbell Plateau, and 60 to 70% of the total U for the cores located in the Solander Trough region. The detrital Re corrections were less than 1% of the total Re for the cores located on the Campbell Plateau and up to 8% for the cores in the Solander Trough region.

3. Results

140 In the following sections the LGM and Holocene time slices have been defined as intervals ranging between 25 to 18 ka and 12 to 5 ka, respectively.

3.1. Authigenic Uranium and Rhenium

The sedimentary aU and aRe concentrations were higher during the LGM compared to the Holocene in the cores located on the Campbell and Challenger plateaux (Figure
145 3). On the Campbell Plateau, the aU concentrations in V1439 were 4.5 mg/kg higher

during the LGM compared to the Holocene. However, for the other sites on the Campbell Plateau (Y16, Y17, SO136GC-38, Y9), the aU concentrations were only approximately 3 mg/kg higher during the LGM. For the same cores aRe concentrations were generally 10 to 20 $\mu\text{g/kg}$ higher during the LGM. On the Challenger Plateau (cores MD06-2988 and DSDP 593) the aU and aRe concentrations were on average 2 mg/kg and 7.5 $\mu\text{g/kg}$ higher respectively, during the LGM relative to the Holocene. Overall, larger aU and aRe concentration differences are observed between the LGM and the Holocene on the Campbell Plateau compared to the Challenger Plateau.

Figure 3: Difference in authigenic U and Re (aU: dots and aRe: diamonds) between the LGM (18-25 ka) and the Holocene (5-12 ka). Positive values indicate higher aU and aRe concentrations in the sediment during the LGM, and lower inferred bottom water oxygen concentrations.

Q217 as well as the cores located in the Solander Trough, which sits between the

Macquarie ridge and the Campbell Plateau (TAN1106-11, TAN1106-07, TAN0803-09) had slightly lower aU and aRe concentrations during the LGM compared to the Holocene. However, the differences for these five cores were modest: aU and aRe differences between the LGM and the Holocene never exceeded 1.5 mg/kg and 5 $\mu\text{g/kg}$ respectively. Y8 did not contain any aU during the LGM or the Holocene.

3.2. Benthic $\delta^{13}\text{C}$

To help determine whether the inferred oxygen changes were primarily controlled by changes in ocean circulation the existing benthic carbon isotope data for this region were compiled (both published and unpublished). The benthic foraminifera $\delta^{13}\text{C}$ values observed for the Challenger and Campbell plateaux during the Holocene agree well with the modern Dissolved Inorganic Carbon (DIC) values (Figure 1 and 4).

Figure 4: Authigenic U and Re (aU and aRe) and benthic foraminifera $\delta^{13}\text{C}$ as a function of core depth for LGM (dashed; 18-25 ka) and Holocene (plain; 5-12 ka) time slices on the Challenger (blue triangles) and Campbell (green circles) plateaux. Standard errors are reported for aU and aRe corresponding to the method precision (See Durand et al. (2016)). Y8 and the Solander Trough sites were excluded because of the potential "burn-down" affecting Y8 and EP changes affecting the Solander Trough sites.

The benthic foraminifera $\delta^{13}C$ variations with depth show a dramatic regional difference between the Campbell and Challenger plateaux (Figure 4). On the Challenger Plateau, similar $\delta^{13}C$ values are observed during the LGM and the Holocene at all depths. However, on the Campbell Plateau, below 600 m, cores show $\delta^{13}C$ increases (0.4-0.85‰) between the LGM and the Holocene that are larger than the whole-ocean glacial-interglacial $\delta^{13}C$ change (0.34‰, Figure 4) (Peterson et al., 2014). Cores on the Campbell Plateau above 600m (MD97-2109, SO13-061 and Y17) show no $\delta^{13}C$ change between the LGM and the Holocene (Figure 4).

4. Discussion

At typical AAIW depths (500-1200 m) on the Campbell and Challenger plateaux, the aU and aRe concentrations were higher during the LGM than in sediments presently bathed by UCDW (1500-2500 m). In contrast, waters of the Solander Trough region (where TAN1106-07, TAN1106-11 and TAN0803-9 were retrieved) were depleted in aU and aRe during the LGM compared to the Holocene. If the observed changes in redox sensitive metal concentrations are interpreted in terms of oxygenation changes at the water-sediment interface, these data suggest the intermediate depths above the Campbell and Challenger plateaux were depleted in oxygen during the LGM compared to the Holocene. As several other factors may drive changes in authigenic metal (aM: U and Re) concentrations, we address each of these in turn.

Sedimentation rate changes can influence the incorporation of aM in the sediments independently of any changes in bottom water oxygenation changes by diluting (increased sedimentation) or concentrating (decreased sedimentation) the aM (Tribovillard et al., 2006). Table 3 compiles sedimentation rates at different sites during the Holocene and the LGM. Overall, no coherent pattern of sedimentation rate changes between the LGM and the Holocene can explain the aU and aRe glacial enrichments observed in the New Zealand region. Three cores had moderately higher sedimentation rates during the LGM compared to the Holocene. Three cores had lower sedimentation rates during the LGM compared to the Holocene and two cores had no change. Only TAN0803-09 and TAN1106-11 had noticeably higher sedimentation rates during the

195 LGM compared to the Holocene. However, these two cores had higher aU and aRe concentrations during the Holocene compared to the LGM. Therefore, we conclude that the aU and aRe variations observed in this study are not primarily controlled by changes in sediment accumulation.

Table 3: Sedimentation rates during the Holocene and the LGM at different core sites. MD06-2988 and Q217 sedimentation rates were not calculated due to insufficient radiocarbon ages.

Core	Holocene sed.rate (cm/ka)	LGM sed.rate (cm/ka)
DSDP 593	3.3	3.3
SO136GC-38	1.7	3.3
TAN0803-09	4.9	19.8
TAN1106-07	7.4	8.7
TAN1106-11	5.7	16.2
V1439	2.9	3.9
Y8	3.9	3.9
Y9	6	1.2
Y16	7.2	1
Y17	7.5	1.5

200 Similarly, drastic changes in sedimentary redox conditions can lead to post-depositional remobilization of U and Re ("burn-down"), potentially complicating paleoceanographic interpretations (Crusius and Thomson, 2000, 2003). Particularly, because most sites in this study have low sedimentation rates (Table 3) they can be affected by remobilisation. To limit the impact of remobilisation on aU and aRe, time slices were defined across large time intervals. However, for Y8, the absence of aU across all samples suggests
205 that this site may have experienced a "burn-down" of aU deposited during the LGM. This "burn-down" would have occurred because of the deepening of the AAIW-UCDW boundary between the LGM and the Holocene (see section 4.2). At other sites, we suggest that oxygenation changes have driven the aU and aRe variations observed. The coherent spatial and temporal aM variations between the sites despite differences in

210 sedimentation rates, show that remobilisation is not a dominant driver of aM variations. Furthermore, the range of aU and aRe concentrations observed in the sediments from our study is consistent with what has been reported previously in the rest of the Southern Ocean with concentrations ranging from 0 to 5 mg/kg and 0 to 20 $\mu\text{g/kg}$ for aU and aRe respectively (Chase et al., 2001; Muratli et al., 2009; Chase et al., 2003; Hayes
215 et al., 2014; Gottschalk et al., 2016; Jaccard et al., 2016). Finally, the similar behaviour between aU and aRe enrichments increases the confidence that oxygen changes drove the aM concentration variations at all sites except Y8 (Figures 3 and 4).

The following sections discuss the mechanisms that could possibly have driven the substantial oxygen changes inferred from aU and aRe variations between the LGM and
220 the Holocene.

4.1. Changes in export production as potential drivers of oxygenation changes

Since labile organic matter is supplied by export production (EP), an increase in EP leads to oxygen depletion in the ocean interior. As a consequence, increased EP in the New Zealand region during the LGM could potentially explain aU and aRe
225 enrichments in the sediments of the Challenger and Campbell plateaux. Durand et al. (2017) used ^{230}Th -Thorium normalised biogenic fluxes from four cores used in this study (DSDP 593, TAN0803-09, SO136GC-38 and Y9) to reconstruct EP changes in the New Zealand region since the LGM. They showed that EP stayed low and rather constant over the Campbell and Challenger plateaux since the LGM. This result is in
230 agreement with previous findings (Ellwood et al., 2005) and rules out increased organic matter remineralisation as a major driver of oxygen changes between the LGM and the Holocene over the Campbell and Challenger plateaux. However, at site TAN0803-09, Durand et al. (2017) found that EP was greater during the deglaciation and Holocene compared to the LGM. This was likely caused by a southward shift of the highly
235 productive subtropical front over the core location during the deglaciation (12-18 ka) (Bostock et al., 2015). Therefore, we suggest that the high aU and aRe values observed during the Holocene for the three Solander Trough cores may have been influenced by increased EP due to changes in ocean circulation.

4.2. Shallower AAIW over the Campbell Plateau during the LGM

240 By plotting the temporal aU and aRe variations along depth, it is apparent that depth and thus the water mass geometry has been a driver of change on the Campbell Plateau between the LGM and the Holocene (Figure 4, S1). On the Campbell and Challenger plateaux, at intermediate depths (500-1200 m), the aU and aRe concentrations were higher during the LGM than in sediments presently bathed by UCDW (1500-2500 m).
245 The higher oxygen concentrations inferred at intermediate depths over the Campbell Plateau during the Holocene compared to the LGM could be partly explained by a deepening of the AAIW-UCDW boundary between the LGM and the Holocene. And if this deepening was associated with a lower UCDW oxygen content during the LGM compared to the Holocene. Decrease in ventilation and thus subsequent reduction in oxygen
250 content of UCDW had been reported previously in the New Zealand region (Skinner et al., 2015). During the Holocene, this deepening would have brought oxygenated AAIW to sites that were bathed by oxygen-depleted UCDW during the LGM. Such changes could help explain the aU and aRe variations at core sites V1439, SO136G-C38 and Y9, Y16 and Y17. Similarly, deepening the UCDW-LCDW boundary between
255 the LGM and the Holocene could explain the higher aM concentrations observed during the Holocene at deeper site, Q217, because UCDW has a lower oxygen concentration than LCDW (Figure 2).

Several studies using benthic foraminifera $\delta^{13}C$ variations in sediment cores concluded that the AAIW-UCDW boundary deepened between the LGM and the Holocene
260 in the Pacific sector of the Southern Ocean (Pahnke and Zahn, 2005; Ronge et al., 2015). These conclusions are based on observations of lower benthic foraminifera $\delta^{13}C$ values during the LGM compared to the Holocene in MD97-2120 (Figure 4). Even though several factors such as differences in biological productivity or air-sea gas exchange can compromise the use of $\delta^{13}C$ as a circulation proxy (Schmittner et al., 2013, 2017), in
265 this region, these studies concluded that circulation is the main driver of $\delta^{13}C$ changes (Ronge et al., 2015; Pahnke and Zahn, 2005). Therefore, because the $\delta^{13}C$ of AAIW ($\approx 1.5\text{‰}$) is higher than that of UCDW ($< 0.5\text{‰}$), these studies concluded that UCDW and not AAIW was bathing the location of MD97-2120 (1210 m) during the LGM.

The benthic $\delta^{13}C$ data from the Campbell Plateau compiled here agree with the

findings from Pahnke and Zahn (2005) and Ronge et al. (2015) and suggest that AAIW was shallower during the LGM compared to the Holocene (Figure 4). The $\delta^{13}\text{C}$ increases observed at V1439 and MD97-2120 between the LGM and the Holocene cannot only be driven by the glacial-interglacial reservoir change but can be explained by the deepening of the AAIW-UCDW boundary between the LGM and the Holocene. However, even though SO213-84-1 is deeper than V1439, it observed a smaller $\delta^{13}\text{C}$ change between the LGM and Holocene (Figure S2). This change is equivalent to the glacial-interglacial reservoir change and suggests that AAIW overlaid this site since the LGM without interruption. This paradox can be explained because SO213-84-1 is located almost 10° further north compared to V1439 and AAIW extends deeper as it flows northward. Presently, the well oxygenated core of AAIW extends to 500 m deep at 60°S and extends to 1500 m at 50°S (Figure 2). Therefore, we suggest that during the LGM the AAIW-UCDW boundary resided above 790 m at 53°S and between 900 and 1200 m at 45°S . This means that at 53°S , AAIW presently reaches depths almost twice as deep as during the LGM.

A shallower AAIW during the LGM above the Campbell Plateau can be explained by a decrease in the AAIW formation rate (Figure 5). A decrease in the AAIW production could have resulted from a northward shift of the westerlies towards lower latitudes (Voigt et al., 2015; Downes et al., 2011), stronger Split Jet or a negative Southern Annular Mode (SAM) (Chiang et al., 2018). Some authors also suggested that during the LGM, increased sea-ice melt would have increased buoyancy forcing (Pahnke and Zahn, 2005), thereby decreasing AAIW density, limiting its subduction depth (Ronge et al., 2015).

4.3. Change in the AAIW source reaching the Challenger Plateau

On the Campbell Plateau both the higher concentrations of aU and aRe and lower $\delta^{13}\text{C}$ during the LGM suggest that AAIW was deeper during the Holocene compared to the LGM (Figures 3, 4). However, the single mechanism of deepening the AAIW-UCDW boundary between the LGM and the Holocene cannot explain all of the data presented in this study. On the Challenger Plateau, aM data show lower oxygenation during the LGM relative to the Holocene, as on the Campbell Plateau. However, in

Figure 5: Schematics of the scenario that can explain the differences between the Campbell (1) and Challenger (2) plateaux $\delta^{13}C$ and redox-sensitive metal data. On the top panel, modern circulation. On the bottom panel the southern source AAIW (red) shoaled and left the Campbell Plateau partly bathed by oxygen depleted UCDW, while the lower oxygen northern source AAIW (yellow) intensified and bathed the Challenger Plateau and shallow parts of the Campbell Plateau.

300 contrast to the cores on the Campbell Plateau, our compilation of $\delta^{13}C$ data from cores DSDP 593, MD06-2990, SO136-003 and MD06-2986 on the Challenger Plateau shows very little difference in $\delta^{13}C$ between the LGM and the Holocene (Figure 4). Indeed, if anything, after the application of a whole-ocean $\delta^{13}C$ change (Figure S2), $\delta^{13}C$ on the Challenger plateau at intermediate depths is slightly higher during the LGM
 305 relative to the Holocene. These data are inconsistent with the deepening of AAIW-UCDW boundary between the LGM and the Holocene, discussed previously for the Campbell Plateau. Instead, the $\delta^{13}C$ data suggest that AAIW uninterruptedly overlaid the Challenger Plateau since the LGM.

A change in the AAIW source bathing the New Zealand region could explain the
 310 oxygen changes on the Challenger Plateau between the LGM and the Holocene (Figure
 5). As mentioned earlier, there are three types of AAIW present around New Zealand
 (Figure 1) (Bostock et al., 2013). In terms of oxygen concentrations, the southern
 sourced AAIW is younger and has higher oxygen content than the northern types
 (Chiswell et al., 2015); however, in terms of $\delta^{13}C$ the three AAIW types have similar
 315 signatures (Figure 2). Presently the southern type AAIW is bathing the Challenger
 and Campbell plateaux (Chiswell et al., 2015). However, it is possible that during the
 LGM the region was bathed by and a northern source AAIW. Muratli et al. (2009)
 argued that the production of AAIW in the southeast Pacific was higher during the
 LGM. The southeast Pacific AAIW is the source of the northern types AAIW, both
 320 present around New Zealand (Figure 1). Therefore, increased AAIW production in the
 southeast Pacific during the LGM could have resulted in the expansion of the northern
 types of AAIW in the New Zealand region. In this case, expanded northern type AAIW
 could have bathed the Challenger Plateau (Figure 5).

4.4. *Implications and perspectives*

325 The glacial deoxygenation observed here is at odds with the general notion that
 AAIW ventilation at intermediate depths was greater during the LGM in the Southern
 Ocean (Jaccard et al., 2014). This increased ventilation is thought to have resulted
 from colder temperatures, which increase the oxygen content of AAIW, and stronger
 westerly winds that are thought to enhance AAIW formation (Downes et al., 2009,
 330 2010). Moreover, the results from this study are also in opposition to those of Muratli
 et al. (2009) who, using redox sensitive proxies, reported higher oxygen content at
 intermediate depths in the southeast Pacific sector of the Southern Ocean (along the
 Chilean Margin). Therefore, these opposing patterns stress the complex responses of
 AAIW ventilation to climate forcing. In particular, the results presented here highlight
 335 an asymmetry of oxygenation at intermediate depths between the eastern and western
 sides of the Pacific sector of the Southern Ocean. A recent study revealed a similar
 oxygenation asymmetry associated with weaker Split Jet and positive SAM (Chiang
 et al., 2018). In their study, they observed increased oxygenation in the South Pacific

globally, linked with increased AAIW formation during weaker Split Jet and/or positive SAM, except in a small zone along the Chilean Margin where the cores studied by Muratli et al. (2009) are located. This deoxygenation along the Chilean Margin originates from the North and is associated with surface undercurrent and PDW and not from a reduced AAIW formation (Chiang et al., 2018). Therefore, it is possible that weaker Split Jet and/or positive SAM developed since the LGM, resulting in increased AAIW formation and increased oxygenation in the South Pacific globally except in a small zone along the Chilean margin where an opposite pattern emerged. Future work will need to address this discrepancy. New redox-metal measurements from multiple sediment cores from the southeast Pacific, spanning a wider range of locations and depths, could help to solve this question. So far AAIW oxygenation reconstruction in this region is based on a limited record (three sites) spanning a small depth range (≈ 500 m) (Muratli et al., 2009). Being able to reconstruct AAIW ventilation and circulation is critical because AAIW plays an important role in sequestering greenhouse gas in the ocean interior (Sabine et al., 2004). Therefore, the reliability of future climate change scenarios depends on our ability to resolve how AAIW responds to climatic forcing.

5. Conclusion

In summary, this study shows that the intermediate depths of the Campbell and Challenger plateaux were less oxygenated during the LGM compared to the Holocene, while in the Solander Trough region no oxygenation change was observed between the LGM and Holocene. In the Solander Trough region this was likely partially caused by an increase in EP from the shifting of the subtropical front over the region during the deglaciation. On the Challenger Plateau, $\delta^{13}\text{C}$ variations show that AAIW continued overlaying the plateau. Therefore, we hypothesise that a change in the dominant AAIW source, from a southern source AAIW to a lower oxygen, northern source AAIW can explain these results. In contrast, a relatively shallower AAIW over the Campbell Plateau during the LGM compared to the Holocene, left deeper parts of the plateau bathed by the low oxygen UCDW. Finally, aU and aRe data suggest that UCDW oxygen content was lower during the LGM than it is presently.

Acknowledgements

This work was supported by the Australia/New Zealand IODP Consortium (ANZIC),
370 and by an ARC Future Fellowship awarded to Zanna Chase [grant number FT120100759].
Access to ICP-MS instrumentation was supported through ARC LIEF funds [grant
number LE0989539]. Samuel L. Jaccard was funded by the Swiss National Foundation
(grant PP00P2-144811). Radiocarbon analyses at ANSTO were supported by a AINSE
grants ALNGRA11081 and ALNGRA14503. Helen Bostock is supported by COPR
375 funding from NIWA (New Zealand government). Thanks to IODP Kochi Core Center
for access to the samples for DSDP 593.

Database

Pangaea database compiling the new data created during this work, including sample
intervals and age models:

380 <https://doi.pangaea.de/10.1594/PANGAEA.864666>

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Supplementary figures

Figure S1: Differences in authigenic U and Re between the LGM (18-25 ka) and the Holocene (5-12 ka) as a function of depth and latitude. Bathymetric transect at 182° was plotted.

Figure S2: Same as Figure 4 except the glacial $\delta^{13}C$ values were corrected for the global isotopic shift by adding 0.34‰ (Peterson et al., 2014).