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Deep sea corals off Brazil verify a poorly ventilated Southern Pacific Ocean during H2, H1 and the Younger Dryas

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ABSTRACT

Simultaneous ^{14}C and Th/U dating of deep sea corals are useful for reconstructing the intensity of deep ocean circulation in the past, as they deliver the time between the gas exchange of the water with the atmosphere and the incorporation of the ^{14}C in the carbonates (Adkins and Boyle, 1997; Adkins et al., 1998; Mangini et al., 1998). Th/U ages of deep sea corals sampled in sediment cores from locations off the coast of Brazil bathed by Antarctic Intermediate Water at depths between 600 and 800 m group close to Heinrich events H2, H1 and the Younger Dryas. The $\Delta^{14}C$ of the water bathing the corals starts to decrease approximately 2 kyr before the Heinrich events and decreases to values 400‰ lower than the corresponding back tracked atmospheric values. The timing and the magnitude of the decrease is similar to that observed in intermediate water in the N. Pacific off Baja California (Marchitto et al., 2007) and in the Eastern Pacific (Stott et al., 2009). High ventilation ages, partly exceeding 4000 years, are an unambiguous indication for a reduction of North Atlantic deep water formation during H2, H1 and the YD, as deduced from higher $^{231}Pa/^{230}Th$ activity ratios and from εNd in N. Atlantic Ocean sediments (McManus et al., 2004; Pahnke et al., 2008; Yu et al., 1996). They also could indicate a poorly oxygenated Southern Pacific Ocean at the end of the Heinrich events.

1. Introduction

The strength of the Atlantic Meridional Overturning Circulation is believed to have crucial influence on global climate. Several different proxies, such as $^{231}Pa/^{230}Th$ activity ratios in sediments and εNd in the post depositional Fe–Mn coatings of sediments have been applied for reconstruction of the deep water structure and strength (McManus et al., 2004; Pahnke et al., 2008; Yu et al., 1996). Deep sea corals provide another high resolution archive of ocean circulation as they record calendar age (U/Th ages) and ^{14}C concentration of the surrounding water masses (Mangini et al. 1998; Adkins et al. 1998). Accordingly, they are a direct archive for atmospheric ^{14}C variations, ageing of water masses and hence circulation changes. Robinson et al. (2005) detected pronounced switches between radiocarbon-enriched and depleted waters during the deglaciation in the Western North Atlantic. The available data support the concept of reduced NADW export during H1 and probably earlier Heinrich Events as well. However, this concept urgently needs validation.

2. Sample locations

We analyzed deep sea corals sampled in sediments taken from off the shore of Brazil, from two locations that are about 400 km apart (Fig. 1). These localities are bathed today by Antarctic Intermediate Water (AAIW). It irrigates the middle slope between 550 m and 1200 m (Viana et al., 1998).

The two sediment cores taken: Piston core ENG-111, hereafter referred to as core C1, Campos Basin continental slope, water depth 621 m; Piston core 21210009, hereafter referred to as core C2, Santos Basin continental slope, water depth 781 m. C1 was sampled on an elongated mud mound with abundant coral bushes and recovered specimens of *Lophelia pertusa*. Core C2 was taken in a pockmark field and recovered both *L. pertusa* and *Solenosmilia variabilis*. In both cores, the sediment consists of hemipelagic mud (marl, carbonate-rich and carbonate-poor mud) with biogenic components consisting mainly of coral fragments and planktonic foraminifera. The carbonate coral mounds in the Campos Basin where C1 was taken were first reported by Viana et al. (Viana et al., 1998). C2 was collected in a pockmark field in

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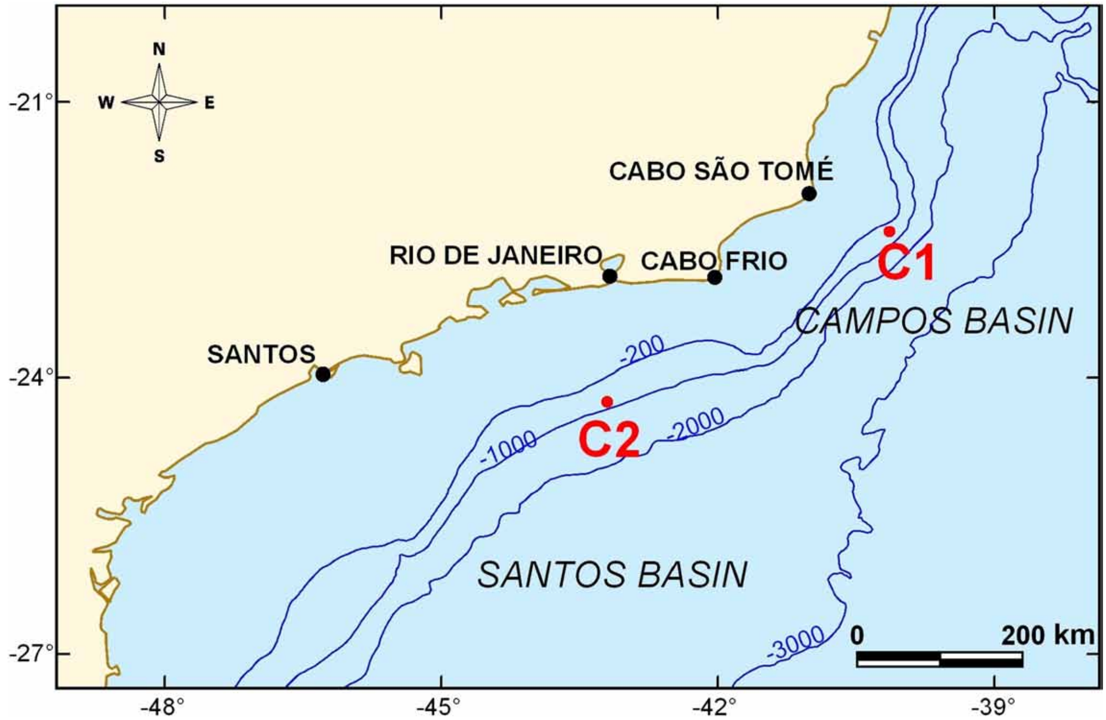

Fig. 1. Location of the sediment cores C1 and C2.

the opposite end of the Santos Basin. The morphology there is similar to the one described by Sumida et al. (2004).

The corals occur throughout about half of the sediment columns for C1 and C2 cores, as shown on Fig. 2. A darker sediment section from 300 cm to the basis of C1 at 380 cm does not display corals. Similarly, core C2 displays no corals in the section between 70 and 220 cm depth.

3. Methods

3.1. Th/U dating of the corals in cores C1 and C2

Coral samples for Th/U dating were taken from approximately 300 cm thick top sections of cores C1 and C2. A first series of samples from core C1 consisted of few mg of materials, as they were the remains of the corals used for ^{14}C dating. These samples have only been cleaned with water and were therefore partly contaminated with ^{230}Th and ^{232}Th adsorbed on their surface. Consequently, they also have larger uncertainties of the ages. They are marked on Table 1 as “old series”. Instead, all the samples from core C2 and a second series from core C1 consisted of larger pieces that had been precleaned with ultrasonic and mechanical procedures, followed by gentle dilution in 0.1 N HNO_3 steps, as described elsewhere (Schröder-Ritzrau et al., 2003).

The sample weights ranged from 0.1 to 0.6 g. The solutions were spiked with a $^{233}U/^{236}U$ double spike and a ^{229}Th spike. Uranium and thorium were isolated by iron co-precipitation and anion exchange columns (Dowex 1X8). The samples were loaded on a preheated rhenium filament and thorium and uranium were measured with a multicollector mass spectrometer (Finnigan MAT 262 RPQ) following instrumental analysis described by Scholz et al. (2004). The Th/U ages of deep sea corals were corrected for initial ^{230}Th, which does not accrue from decay of ^{234}U in the sample, as described by Cheng et al. (2000). The calibration of our spikes is based on the same HU-1 reference material used by N. Frank (Frank et al., 2009). This calibration is in very good agreement with that applied in Bristol (Hoffmann, pers. comm.).

The results derived for the coral samples in sediment cores C1 and C2 are listed in Table 1. Uncorrected and corrected ages are listed in two columns together with their 2 sigma uncertainties, respectively. This correction is performed using the ^{232}Th content of the samples under the assumption of an activity ratio of $^{230}Th/^{232}Th$ in the water column of 8 ± 4 (Frank et al., 2009; Schröder-Ritzrau et al., 2003). The uncertainty of this correction was propagated into the age uncertainty. After the correction five samples of the old series with ^{232}Th contents exceeding a 10 ng/g plot on the depth/age relationship (Fig. 3) close to samples with negligible corrections suggesting that the applied correction factor is approximately right. Nevertheless, these five samples with ^{232}Th contents exceeding 10 ng/g were rejected for the determination of $\Delta^{14}C$.

All the initial $\delta^{234}U$ values of corals from C1 and C2 in the section between 8000 and 15,000 years (Fig. 4) are within the expected range for the present day sea water of 146.6–149.6‰ (Delanghe et al., 2002 ; Robinson et al., 2004). The average of these samples, 145.5 ± 3.4‰, comes close to the average initial $\delta^{234}U$ of 148 ± 3.5‰ observed in cogenetic corals in the N. Atlantic (Frank et al., 2009). In contrast, the corals in the section between 15,000 years and 20,000 years display a slightly lower initial $\delta^{234}U$ of 139.9 ± 2.9‰ than the samples in the younger section. These lower values could be interpreted either as due to diagenetic alteration of the samples or to a lower $\delta^{234}U$ of glacial sea water. We exclude diagenetic processes as the cause of the lower glacial values in corals off Brazil because glacial reef corals off Barbados and Papua also display lower $\delta^{234}U$ values than younger ones, as compiled by Esat (Esat and Yokoyama, 2006). The lower glacial values in the corals off Brazil support their conclusion of variable uranium isotope composition in the ocean over glacial-interglacial scales. In summary, except for two samples from core C2 at depths below 220 cm, corresponding to the last Interglacial period, the initial $\delta^{234}U$ shows no increase with age, as one would expect for older samples being exposed for longer time to exchange with sea water uranium (Neff et al., 1999).

3.2. ^{14}C dating and derivation of $\Delta^{14}C$ in cores C1 and C2

The ^{14}C dating of coral samples was performed at the Keck-CCAMS facility, USA, and at the ^{14}C AMS Lab of the ETH-Zurich. Both labs require graphite samples to perform ^{14}C measurements on their Accelerator

Fig. 2. (a and b): Photographs of the two sediment cores from the Campos (C1) and Santos Regions (C2). Corals in the upper 300 cm of core C1 are dated between 8000 yr and 20,600 years. In C2 reliable Th/U ages between 10,500 and 29,300 years could only be derived in the 70 cm long upper section. Corals from the section between 220 cm and 363 cm of core C2 were diagenetically altered and could not be dated with Th/U. The 20 and 30 cm gap in the cores was due to onboard sampling unrelated to this study.

Mass Spectrometry (AMS) systems. To achieve this requirement using carbonaceous starting materials, carbon was chemically separated from the original sample by CO_2 production (from a chemically clean sample) and then converted to graphite target.

CO_2 was produced from 50% pre-leached coral samples by acid hydrolysis using 85% phosphoric acid. The CO_2 produced was cryogenically purified inside a vacuum line reduced to graphite (Santos et al., 2004; Unkel et al., 2007).

The graphite samples were measured either on the compact AMS (NEC 0.5MV 1.5SDH-2) at Keck-CCAMS (Southon et al., 2004) and at the Mini Carbon Dating System at ETH-Zurich (Synal et al., 2004). The individual targets were counted for 50,000 ^{14}C events at Keck-CCAMS and for 10,000–30,000 at ETH-Zurich (Table 2). The errors were calculated based on counting statistics and scatter in multiple measurements on each sample, along with propagated uncertainties from normalization to standards, background subtraction (based on measurements of ^{14}C free material), and isotopic fractionation corrections, following instrumental analysis described in (Santos et al., 2007; Wacker et al., 2009).

The values $\Delta^{14}C$ (in ‰) for the corals from sediment cores C1 and C2 were derived as:

$$\Delta^{14}C \;=\; \left(\left(\exp\left(-{}^{14}C_{\mathrm{age}}/8033\right) \;/\; \exp\left(-Th/U_{\mathrm{age}}/8266\right)\right)-1\right)*1000$$

as described in (Adkins and Boyle, 1997). These values together with their 1 sigma uncertainties are listed in Table 2.

The $\delta^{13}C$ values were measured on the prepared graphite using the AMS spectrometer. These can differ slightly (typically 1–3‰ or less) from those of original material, if isotopic fractionation occurred during sample graphitization and/or the AMS measurement.

3.3. Trace element analyses (IRD)

About 1.0 gram sample was cleaned as described by Lomitschka and Mangini (Lomitschka et al., 1997) and grounded. After that, circa of 0.25 g was taken and dissolved with 25 ml supra-pure 3 M HNO_3. One ml aliquot was taken, diluted to 1:10 with water quality Milli-Q® and Cd determined by ICP-MS applying In and Tl as internal standard (Godoy et al., 2004). The same solution was applied for the determination of Ca using ICP-OES.

The Cd/Ca ratios derived for the samples from core C1 are listed in Table 1.

4. Results

4.1. The growth periods of the coral samples

Most of the coral samples from the younger sediment sections of C1 and C2 gather around the Heinrich events H2 and H1 and their precursors and the Younger Dryas (Bard et al., 2000; Pahnke et al., 2008) (Fig. 3). The corals from the uppermost 300 cm of sediment core C1 have ages younger than 20.6 kyr and coral growth follows H1 and a precursor event of H1 with low SST as observed in sediment cores off Brazil (Jaeschke et al., 2007). The ages of the corals in this section of the sediment deliver an apparent sedimentation rate of approximately 40 cm/1000 years. However, this estimate of the sedimentation rate has a larger uncertainty, as the coral fragments in the sediment cores may originate from different coral patches allowing even for the inversion of ages down core. For example, the Th/U ages of two different coral samples taken from the same depth from core C2 (44 cm Table 1) differ by 5800 years. This discrepancy is replicated by the ^{14}C dating delivering a similar offset of the ages (Table 2).

Corals from a depth between 5 and 70 cm in C2 have ages younger than 29.3 kyr. They follow the H3 event, dated between 31 and 28 kyr (Veiga-Pires and Hillaire-Marcel, 1999) as well as a precursor event of H2 at 27 kyr (Jaeschke et al., 2007).

Two corals from the sections below the grey interval (at 220 cm) in C2, have very low $\delta^{234}U$ (20 and 15‰), suggesting post depositional diagenetic alteration. The planktonic foraminifera zonation (Ericson and Wollin, 1968) places these core sections within the X zone, encompassing the Last Interglacial Period (M. A. Vicalvi, personal communication, 2008). These two samples were not used for dating and are not listed in Table 1.

4.2. The $\Delta^{14}C$ of deep water

The simultaneous determination of ^{14}C and Th/U ages on one coral sample delivers the $\Delta^{14}C$ of deep water. This value is then "tracked back", correcting for radioactive decay of ^{14}C to the atmospheric value in the past, to derive the ventilation ages, i.e. the time since the water mass last exchanged with the atmospheric ^{14}C (Adkins and Boyle, 1997; Adkins et al., 1998; Mangini et al., 1998).

The comparison of the deep water $\Delta^{14}C$ derived from the corals from both cores C1 and C2 with the curve of atmospheric $\Delta^{14}C$ (IntCal-09) (Reimer et al., 2009) in Fig. 5 shows that several samples are close to the atmospheric $\Delta^{14}C$ values, as expected for corals bathed by South Atlantic Central Water (SACW), which is in closer contact with upper water mass.

However, the samples from C2 and C1 that grew during H2, H1 and the YD have been bathed in water significantly depleted in ^{14}C. A lowering of deep water radiocarbon by as much as to 200‰ compared

Table 1
Results of Th and U measurements on the corals from cores C1 and C2. The sample weights ranged from 0.1 to 0.6 g. Except for the samples with few mg of materials from core C1 (old series). Corrected ages are those derived after a correction for the initial ^{230}Th. This correction is performed using the ^{232}Th content of the samples under the assumption of an activity ratio of $^{230}Th/^{232}Th$ in the water column of 8 ± 4. The Cd/Ca ratios derived for the samples from core C1 are listed in column 15.

Description	Depth (cm)	$\delta^{234}U$ (initial) (‰)	Error 2σ (‰)	^{238}U (µg/g)	Error 2σ	^{232}Th (ng/g)	Error 2σ	^{230}Th (pg/g)	Error 2σ	Age uncorr. (yr BP)	Error 2σ (yr)	Corr. age. (yr BP)	Error 2σ (yr)	Cd/Ca (µmol/mol)
Core C1														
Old series	7	141.9	17.5	3.074	0.003	73.1	1.0	7.1	0.2	14,307	405	8,202	3,685	0.32
Old series	17	140.6	15.0	4.454	0.004	98.2	1.2	10.3	0.4	14,375	197	8,712	3,210	0.21
Old series	32	144.8	3.7	4.533	0.005	3.00	0.02	6.5	0.1	8,599	181	8,442	290	0.31
Old series	52	150.7	4.4	3.364	0.003	4.73	0.02	5.9	0.1	10,714	197	10,372	405	0.11
	54	145.9	2.3	2.765	0.003	0.327	0.002	5.29	0.05	11,717	111	11,692	150	
Old series	70	150.5	7.2	3.17	0.003	2.85	0.03	7.80	0.3	15,336	704	15,112	910	
	113	143.5	4.1	2.767	0.005	1.75	0.02	6.5	0.1	14,544	305	14,392	425	
Old series	130	149.3	5.6	3.301	0.003	16.38	0.07	8.3	0.1	15,599	261	14,362	930	0.05
Old series	152	138.2	4.7	2.759	0.003	6.82	0.05	7.2	0.1	16,521	254	15,902	620	0.23
	163	144.6	2.8	2.913	0.003	2.440	0.008	7.75	0.05	16,723	115	16,522	255	
	193	138.2	2.1	3.629	0.004	0.674	0.002	9.76	0.05	17,028	105	16,982	155	
Old series	230	136.1	7.4	4.100	0.004	7.4	0.1	11.8	0.4	18,447	651	18,002	980	0.15
	235	138.4	2.8	3.304	0.003	1.195	0.006	9.3	0.1	17,866	238	17,782	320	
	247	138.4	3.1	2.975	0.004	1.197	0.006	8.26	0.08	17,617	197	17,522	290	
Old series	262	135.8	8.3	3.340	0.004	15.7	0.1	9.5	0.3	18,178	575	16,992	1,280	0.21
	295	137.0	2.9	3.172	0.003	1.351	0.006	9.39	0.09	18,923	210	18,822	310	
Old series	297	140.4	13.6	3.302	0.003	22.1	0.6	10.9	0.8	21,181	1,677	19,502	2,655	0.1
Core C2														
	5	147.9	2.2	3.932	0.004	5.95	0.02	8.3	0.1	12,990	127	12,612	335	0.2
	17	146.1	3.7	3.765	0.004	1.99	0.02	8.0	0.1	13,058	241	12,932	345	
	30	145.1	1.9	3.797	0.004	2.34	0.01	6.7	0.1	10,684	97	10,532	190	0.19
	44	140.3	2.7	3.659	0.004	0.499	0.003	6.3	0.1	10,555	125	10,522	165	
	44	142.6	3.6	3.674	0.004	1.06	0.01	9.6	0.1	16,422	270	16,352	350	0.05
	47	143.4	2.6	3.871	0.004	5.10	0.02	9.60	0.08	15,512	137	15,182	330	
	55	134.9	5.4	3.354	0.005	4.50	0.05	13.0	0.2	25,573	422	25,242	685	
	58	151.4	4.0	3.017	0.003	2.47	0.02	12.8	0.2	28,072	452	27,872	625	
	59	146.8	2.9	3.120	0.003	1.21	0.01	12.9	0.1	27,165	295	27,072	395	
	63	141.7	2.7	3.062	0.003	0.96	0.01	12.1	0.1	26,030	283	25,952	365	
	65	146.3	2.5	3.507	0.004	0.560	0.003	14.4	0.1	26,972	198	26,932	265	
	68	150.9	2.9	3.775	0.004	3.00	0.01	13.8	0.1	23,713	229	23,522	375	
	68	142.8	4.6	3.269	0.003	0.94	0.01	14.3	0.3	29,340	715	29,272	815	

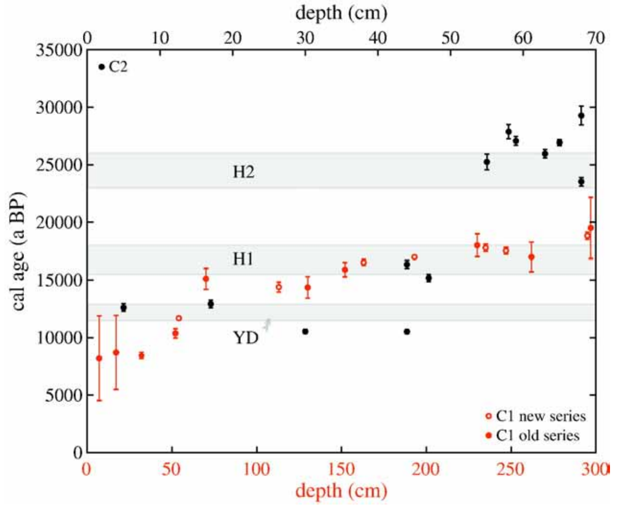

Fig. 3. Th/U ages of the coral samples against depth in the two sediment cores C1 and C2 and corrected for initial ^{230}Th. The horizontal bars mark the timing of the YD, and Heinrich events H1, and H2 (Bard et al., 2000; Pahnke et al., 2008). A first series of samples from the core C1 ("old series") consisted of a few mg of the material, as they were the remains of samples used for the ^{14}C dating. These samples have only been cleaned with water and were therefore partly contaminated with ^{230}Th and ^{232}Th adsorbed on their surface. Consequently, they also have larger uncertainties of the ages.

to the contemporary atmosphere was reconstructed from deep sea coral samples from the N. Atlantic (Robinson et al., 2005). It was attributed to a larger component of SSW in the water column shortly following the Heinrich 1 event and after the Younger Dryas (Robinson et al., 2005). The water bathing the corals off Brazil during H1 is more depleted of ^{14}C by about 200‰ than the N. Atlantic water (Robinson et al., 2005) and less depleted by about 200‰ than the intermediate water in the Eastern North Pacific (Marchitto et al., 2007) (Fig. 5). Another core from the Eastern Equatorial Pacific evidences water that is even more depleted by about 500‰ than in the North Pacific during H1(Stott et al., 2009). The onset of the lowering of $\Delta^{14}C$ off Brazil is contemporaneous to that in the Pacific during H1. At present no highly resolved data from the North Pacific are available in the period

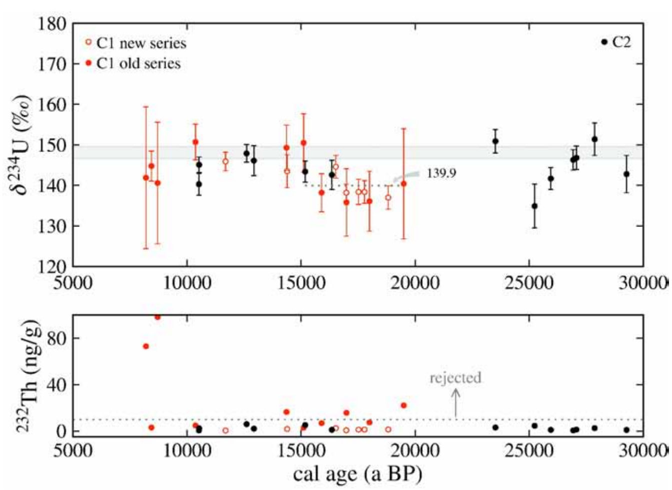

Fig. 4. The initial $\delta^{234}U$ of corals from C1 and C2 in the section between 8000 and 15,100 years are within the expected range for the present day sea water of 146.6–149.6‰ (Delanghe et al., 2002 ; Robinson et al., 2004). The average of these samples (145.5 ± 3.4) ‰ comes very close to the average initial $\delta^{234}U$ 148 ± 3.5‰ observed in cogenetic corals in the N. Atlantic (Frank et al., 2009). In contrast, the corals in the section between 15,100 years and 20,000 years display slightly lower initial $\delta^{234}U$ (139.9 ± 2.8) ‰ than the samples in the younger section.

covering H2. Two data points from Baja California that fall within H2 show no distinct depletion of $\Delta^{14}C$. The large $\Delta^{14}C$ deficiency of the Brazil corals delivers ventilation ages of up to 4000 years compared to the back tracked atmosphere during H2 and H1.

The high ventilation ages off Brazil are not caused by local injection of methane but rather reflect radiocarbon age anomalies in the SSW. Local injections of methane should be accompanied by significantly lower $\delta^{13}C$ values in the corals (Becker et al., 2009; Lietard and Pierre, 2009). Roughly, 1000 years older ^{14}C ages would correspond to approximately 10‰ lighter $\delta^{13}C$. The average $\delta^{13}C$ of the corals measured at the ETHZ of -5.9 ± 1.3‰ ($n = 9$) lies within the average value of -5.9 ± 1‰ ($n = 36$) observed in corals from the Rockall Plateau in the N. Atlantic (Blamart et al., 2005). Furthermore, $\delta^{13}C$ and $\Delta(\Delta^{14}C)$ are not correlated.

4.3. The Cd/Ca ratio of the corals

The Cd/Ca in corals records the ratio in sea water (Adkins et al., 1998). SSW has a larger component of old Pacific water, and thus significantly higher Cd/Ca ratios than the Northern Atlantic water (Broecker, 1995). The concentration of cadmium oscillates between Cd/Ca ratios of 0.05 and 0.3 µmol/mol (Table 2). Although there is yet no calibration showing that these specific coral species record the Cd/Ca ratio of sea water without any alteration, a higher Cd/Ca content of the corals could reflect the intrusion of SSW, in the same manner as lower $\Delta^{14}C$ do. After the correction of the ages for initial ^{230}Th, the Cd/Ca ratio of nine samples of C1 and the 3 samples of core C2 reproduce the pattern reconstructed for $\Delta^{14}C$ as was reported for the N. Atlantic (Robinson et al., 2005), showing a double peak towards older water values following H1. As the corals grew in the uppermost section of SSW the higher Cd/Ca most likely suggest that SSW not only had a lower $\Delta^{14}C$ but that it also had a higher concentration of nutrients during H1.

5. Paleoceanographic implications

Which preliminary conclusions may we derive from these results?

First, $\Delta^{14}C$ values of the water decreasing along the decay curve of radiocarbon and the higher Cd/Ca ratios during H2 and H1 may be explained assuming periodic slower meridional overturning circulation of several thousand years duration as suggested by McManus et al., 2004. The period between 17.5 and 14.5 kyr with a significant lowering of $\Delta^{14}C$ values coincides with the period of maximum sedimentary $^{231}Pa/^{230}Th$ ratios observed in GGC5 off Bermuda (McManus et al., 2004) (Fig. 6). During the period between 23 kyr and 28 kyr the $\Delta^{14}C$ changes dramatically – from zero offset to about 500 per mill – indicating a distinct change during H2. It corresponds to a section with higher $^{231}Pa/^{230}Th$ ratios at Site 1063 during H2, sampled in the vicinity of GGC5 (Lippold et al., 2009). Therefore the coral data are a verification for the predicted reduction/stopping of NADW production derived from the higher $^{231}Pa/^{230}Th$ ratios and from εNd, that were attributed to a stronger influence of the SSW (Pahnke et al., 2008), as well as for the higher ventilation ages observed in N. Atlantic sediments during YD and H1 (Schröder-Ritzrau et al., 2003). The precise ages of the corals off Brazil show that the beginning of progressive aging of the water coincides with periods of significantly lower SST detected north east of Brazil that were attributed to Fennoscandian and Icelandic iceberg discharges preceding H2 and H1 (Jaeschke et al., 2007) (Fig. 6). The beginning of the three periods of slower circulation is accompanied by an increase of atmospheric $\Delta^{14}C$. However, as already discussed by Hughen et al. (1998), a simple shutdown of NADW could explain the increase but not the subsequent drawdown of $\Delta^{14}C$ during the Younger Dryas, implying the presence of additional mechanisms which lower atmospheric $\Delta^{14}C$ during this interval of reduced deep water formation. Subsequent drawdown of atmospheric $\Delta^{14}C$ also occurred during H1 and H2. Possible mechanisms explaining the drawdown

Table 2
Results of ^{14}C measurements on coral samples from cores C1 and C2. Laboratory measurement identifiers: UCIAMS# (from KCCAMS/UCI facility) and ETH# (from ^{14}C AMS Lab of the ETH-Zurich). Column 4: Conventional radiocarbon age (Stuiver and Polach, 1977). Columns 6 and 7: $\Delta^{14}C$ and 1σ error. (Stuiver and Polach, 1977).

Description	Lab#	Depth (cm)	^{14}C age (yr BP)	Error 1σ (yr BP)	$\Delta^{14}C$ (‰)	Error 1σ (‰)	U/Th corr. age (yr BP)	Error 1σ (yr)
Core C1								
Old series	UCIAMS27015	32	10,940	20	−288.65	12.60	8,442	145
Old series	UCIAMS27016	52	11,010	25	−109.36	21.99	10,372	203
	ETH-35227	54	11,137	59	28.46	12.01	11,692	75
	ETH-35225	113	14,135	71	−18.34	26.69	14,392	213
Old series	UCIAMS27019	152	14,540	30	120.47	42.23	15,902	310
	UCIAMS48068	163	14,475	25	217.56	19.16	16,522	128
	ETH-35226	193	15,108	75	189.70	15.74	16,982	78
Old series	UCIAMS27021	230	15,610	35	264.41	75.15	18,002	490
	UCIAMS48066	235	15,555	30	239.66	24.44	17,782	160
	ETH-35229	247	15,661	88	179.05	24.15	17,522	145
Core C2								
	UCIAMS35357	5	10,925	30	180.28	24.32	12,612	168
	UCIAMS35358	17	10,950	35	223.06	26.07	12,932	173
	UCIAMS35359	30	11,135	30	−105.98	10.80	10,532	95
	ETH-35224	44	11,233	53	−117.88	10.55	10,522	83
	UCIAMS35360	44	14,975	40	120.80	24.38	16,352	175
	ETH-35222	55	23,540	180	131.23	53.29	25,242	343
	UCIAMS35362	58	23,000	90	663.12	65.58	27,872	313
	ETH-35221	63	23,260	165	276.41	38.49	25,952	183
	ETH-35220	65	22,710	160	538.92	39.35	26,932	133
	ETH-35223	68	22,950	155	−11.27	29.44	23,522	188

could be enhanced circulation of intermediate water in the Indian Ocean during the Heinrich events (Schulte et al., 1999) or enhanced ventilation of the deep North Pacific (Gebhardt et al., 2008), or both.

In the sections corresponding to the YD and H2 the $\Delta^{14}C$ of the coral samples plot on the decay curve of ^{14}C, reflecting the aging of the source water. In H1 two low ventilation events could be deduced from the data. In the lack of evidence for an increase of atmospheric methane during the deglacial Mystery Interval (Broecker and Barker, 2007), Stott et al. (2009) suggested a scenario based on an intrusion of significantly older N. Pacific intermediate water into the tropical Pacific layers. The higher ventilation ages off Brazil during H1 and the YD, as well as during H2 suggesting that intermediate and deep waters in the Pacific were significantly older prior to H1, are compatible with this scenario. Enhanced overturning in the Atlantic sector of the Southern Ocean during the Mystery Interval has been recently suggested (Anderson et al., 2009). In these scenarios decreasing $\Delta^{14}C$ in intermediate water could be explained by admixing of ^{14}C depleted saline deep water mass that has been isolated from the atmosphere for thousands of years. However, the aging of intermediate water off Brazil along the decay curve of ^{14}C would require very special circumstances to work, because then the balance between deep water renewal and gas exchange rate would need to maintain a drop in $\Delta^{14}C$ similar to the decay rate. We therefore favor the scenario that assumes that the Pacific component of the SSW gradually aged following a shutdown or reduction of the NADW formation. The magnitude and the timing of the decrease of the $\Delta^{14}C$ off Brazil during H1 is analogue to that observed synchronously in intermediate water in the North Pacific (Marchitto et al., 2007) (Fig. 5) and in the eastern Pacific (Stott et al., 2009). A shorter period of restarted ocean circulation around 25 kyr supports this scenario. A slight lowering of the $^{231}Pa/^{230}Th$ ratios in the N. Atlantic, which coincides with warming in the N. Atlantic (Jaeschke et al., 2007) and a cooling of the S. Ocean at around 25 kyr (Barker et al., 2009), was probably not long enough to ventilate the deep Pacific and therefore did not affect the steady increase of ventilation ages off Brazil that were fed with SSW with a component of older deep Pacific water.

In summary, the record of εNd indicates a larger fraction of SSW in the N. Atlantic during YD and H1. Higher $^{231}Pa/^{230}Th$ in the N. Atlantic sediments reflect both, temporarily enhanced scavenging of Pa due to the higher supply of nutrients with the SSW as well as the reduction of transport of Pa towards the S. Atlantic during YD, H1 and H2. The $\Delta^{14}C$ in the corals off Brazil delivers additional information that the Pacific component of the SSW was not ventilated during these periods. Two deep water $\Delta^{14}C$ values from corals from the Drake Passage at 16.7 kyr (Robinson and van de Flierdt, 2009) and at 17.7 kyr (Goldstein et al., 2001) plot on the curve derived off Brazil (Fig. 5), corroborating the common source of the waters. Periods of lower ventilation apparently begin with the precursor events of low salinity.

We reject the scenario where the depletion of $\Delta^{14}C$ during H2, H1 and the YD was caused only by a stagnation of the intermediate water off Brazil, because shallow waters should feel a significant exchange

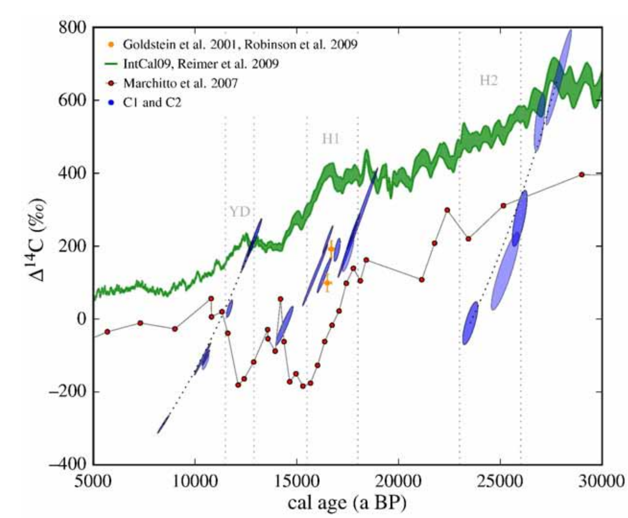

Fig. 5. Comparison of $\Delta^{14}C$ derived from the corals of C1 and C2 with the atmospheric $\Delta^{14}C$, IntCal09 (Reimer et al., 2009). The uncertainties of $\Delta^{14}C$ and of the Th/U ages are plotted as 2σ error ellipses. The corals from both locations show significantly lower $\Delta^{14}C$ values in the sections 23 kyr–26 kyr, 14.5 kyr–17 kyr and 11.7–10.4 kyr, following H2, H1 and the YD. Yellow crosses: two coral samples from the Drake Passage (Goldstein et al., 2001; Robinson and van de Flierdt, 2009). The magnitude and the time behaviour of $\Delta^{14}C$ off Brazil during H1 and the YD is analogue to that observed in intermediate water in the north. Pacific (Marchitto et al., 2007).

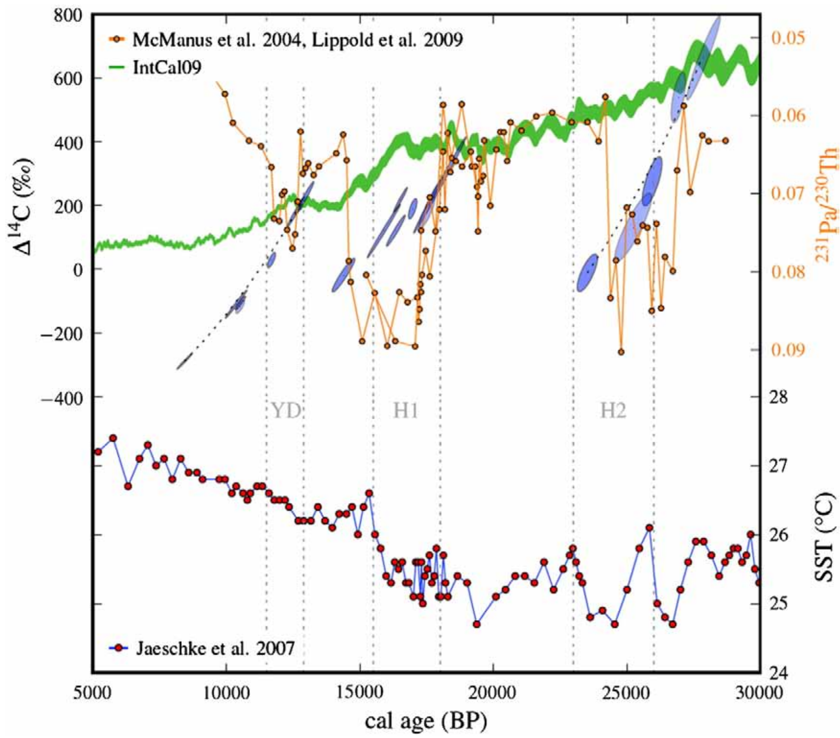

Fig. 6. A comparison of the $\Delta^{14}C$ of the corals off Brazil with the $^{231}Pa/^{230}Th$ ratios in sediment cores from the N. Atlantic. Periods with higher $^{231}Pa/^{230}Th$ ratios in the N. Atlantic coincide with the beginning of the progressive aging of the intermediate water off Brazil. Also plotted are the SST observed in a sediment core off Brazil (Jaeschke et al., 2007).

with the atmosphere. We assume that both cores are bathed in the same SSW water. In fact, the samples from cores C1 and C2 plot close to each other on the decay curves of ^{14}C.

The very high ventilation age at 8.5 kyr is of special interest, because it coincides with a higher ventilation age observed in N. Atlantic corals at 8.9 kyr (Schröder-Ritzrau et al., 2003) and with the age of the Lake Agassiz drainage event at 8470 years (Barber et al., 1999; Kleiven et al., 2008). Does this high ventilation age mark the export of the last remnant of poorly ventilated water from the deep Pacific previous to the beginning of a better ventilation during Late Holocene?

Second, the oxygen content in the deep Pacific Ocean was probably lower than today during periods of low $\Delta^{14}C$. If the oxygen utilization rate (OUR) during these events was similar to the average for the Ocean today, between 0.8 and 1 mmol/m^3‰$^{-1}$ $\Delta(\Delta^{14}C)$ (Broecker et al., 1991) (Sarmiento and Gruber, 2006), deep water $\Delta(\Delta^{14}C)$ values lower than −350‰ than the back tracked atmosphere would then suggest short periods of anoxia of deep water at the end of H1 and H2, explaining the peaks of uranium and manganese observed in the sediments throughout the N. Atlantic and the Pacific Oceans close to H1 (Mangini et al., 1991; Mangini et al., 2001; Mangini et al., 1994). Due to lower $\Delta^{14}C$ values the Pacific Ocean being more susceptible to anoxia than the deep N. Atlantic.

Anoxic periods could also result in the temporarily extinction of benthic species, explaining the lack of similarly old ventilation ages from paired benthic–planktonic records.

Third, we state that coral growth off Brazil was enhanced during three specific periods that coincide with the onset of H2, H1 and YD. The probable explanation is that corals benefited from an enhanced productivity at sea surface during these three periods. However, the process that supplied more nutrients to surface water is unclear. Today the nutrient rich AAIW, as shallow as 500 m, is covered by the South Atlantic Central Water and by Superficial Tropical Water (Viana et al., 1998). Viana et al. (1998) attributed the upwelling of cold SACW to the process of deceleration of the high speed Brazil warm current after passing the Cape of Sao Tome. If this model is correct, then enhanced coral growth would suggest periods of stronger upwelling. Was it due to a N.–S. shift of the Atlantic Intertropical Convergence Zone as speculated by Toledo et al. (2007)? And could the stopping of coral growth and the intermittent growth of deep sea corals off Brazil be attributed to lower oxygen content of the deep water?

Future studies of the $^{231}Pa/^{230}Th$ ratios as well as of trace elements in the coral bearing sediments off Brazil, may help answering some of these questions.

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