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Geochemical evidence of the 8.2 ka event and other Holocene environmental changes recorded in paleolagoon sediments, southeastern Brazil

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ABSTRACT

The paleoclimatic record of Juréia Paleolagoon, coastal southeastern Brazil, includes cyclic and gradual changes with different intensities and frequencies through geological time, and it is controlled by astronomical, geophysical, and geological phenomena. These variations are not due to one single cause, but they result from the interaction of several factors, which act at different temporal and spatial scales. Here, we describe paleoenvironmental evidence regarding climatic and sea level changes from the last 9400 cal yr BP at the Juréia Paleolagoon — one of the main groups of protected South Atlantic ecosystems. Geochemical evidences were used to identify anomalies from multi-proxy analyses of a paleolagoon sediment core. The anomalies of centennial scale were correlated to climate and transgression–regression cycles from the Holocene period. Decadal scale anomalous oscillations in the Quaternary paleolagoon sediments occur between 9400 and 7500 cal yr BP, correlated with long- and short-term natural events, which generated high sedimentation rates, mainly between 8385 and 8375 cal yr BP (10 cm/yr). Our results suggest that a modern-day short-duration North Atlantic climatic event, such as the 8.2 ka event, could affect the environmental equilibrium in South America and intensify the South American Summer Monsoon.

Introduction

Between about 8000 and 5000 years ago (mid-Holocene) significant global climatic changes occurred, and this period is recognized in the Northern Hemisphere as the Holocene Climatic Optimum. Elsewhere, in northern equatorial Africa around 9000 and 7000 cal yr BP, enhanced flood activity occurred along the Nile in association with an increase incursion of rain of Mediterranean origin in the Red Sea (Arz et al., 2003; Kotthoff et al., 2008). Those changes corresponded with a weaker South American Summer Monsoon (SASM) and an apparent shift in the mean position of the Intertropical Convergence Zone (ITCZ) farther north (Cruz et al., 2005; Wang et al., 2005). During this period temperatures were anomalously high in the Northern Hemisphere (Kaufman et al., 2004). Meanwhile, relatively dry climate conditions were identified in southern and southeastern Brazil and along the Chilean coast northward to the Galápagos (Behling et al., 2004; Araujo et al., 2005; Cruz et al., 2006; Ledru et al., 2009). Complex changes in the Peru–Chile coastal circulation indicate lower temperatures than present (−1 to −4°C) and wetter periods in the western Central Andes (Betancourt et al., 2000; Carré et al., in press).

Short-term and abrupt climatic shifts often are difficult to recognize in the sedimentary record. At the end of the last glaciation, global temperatures started to rise in conjunction with the formation of extensive fresh-water lakes that resulted from the melting of large glaciers. Around 8200 yr ago, rapid discharge of glacial meltwater into the North Atlantic Ocean gave rise to changes in global thermohaline circulation (Barber et al., 1999). Known as the 8.2 ka event, this abrupt climatic impact caused the lowering of temperatures for several hundred years in the Northern Hemisphere and changed patterns of human occupation (Gonzalez-Samperiz et al., 2009).

On account of the 8.2 ka event, there was an increase in the relative mean sea level due to the glacial isostatic adjustment, with wide variation in the United States (West and Gulf Coasts) and significantly higher

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signals associated with the catastrophe on the eastern coast of South America and western Africa (Kendall et al., 2008). The Brazilian coast, with an increase of ~1 m of mean sea level, is one of the most suitable places to understand the outflow volume and sea-level signature associated with the 8.2 ka event (Kendall et al., 2008). In the Netherlands, the jump in sea level reached 2.11 ± 0.89 m in 200 yr, which allowed the inference of a global-averaged eustatic sea-level jump of 3.0 ± 1.2 m (Hijma and Cohen, 2010).

Very well-preserved sediments discovered in southeastern Brazil contain a number of environmental indicators that help to understand changes in local sediment dynamics. We herein tentatively associate this sedimentary record with to global-scale climatic and relative sea-level variations. Our study uses a high-resolution Quaternary paleolagoon sediment core with adequate dating to detect minor abrupt sedimentation changes at the Juréia-Itatins Ecological Station (Brazil). In turn, changes in the sedimentary record are correlated with long-term climatic trends and short-term extreme climatic events and sea-level oscillations.

Paleoenvironmental data

Few paleoclimatic and paleoenvironmental data are available in Brazil, with a concentration of data mainly from the Amazon and Cerrado regions. Moreover, the available data often are difficult to comprehend due to differences in the methodologies and data interpretations of different authors and the use of uncalibrated ages.

Several colluviation events have been associated with temperature and vegetal cover reductions and brief periods of heavier precipitation that occurred during paleoclimatic transitions during the last 1 Ma (Sallun and Suguio, 2010). Between about 9000 and 3000 cal yr BP, the climate was drier in South America, according to several records in the Pantanal (Whitney et al., 2011), Andes (Seltzer et al., 1998; Thompson et al., 1998; Baker et al., 2001; Moy et al., 2002; Paduano et al., 2003) and savanna expansion in Amazonia (Servant et al., 1989; Absy et al., 1991). The hydrology of the Amazonian wetlands is strongly influenced by the Atlantic sea level (Behling, 2002), and recent data show that there is an atmospheric connection between South America and the southeastern Pacific (Carré et al., in press). In Northeastern Brazil, it is suggested that semiarid conditions prevailed from 28,000 to 9000 ^{14}C yr BP (~32,500 to 10,000 cal yr BP) and 4535 ^{14}C yr BP (5286–5038 cal yr BP), indicating an expansion of those conditions (De Oliveira et al., 1999).

In the central Amazon (Northern Brazil), drier periods were recorded between 4200 and 550 cal yr BP and between about 6400 and 4600 cal yr BP, which reduced river discharge and caused sea level to fall (Behling and Costa, 2001; Behling, 2002; Irion et al., 2006; De Toledo and Bush, 2008). Cordeiro et al. (2008) identified two drier phases in relation to the present in the Amazon area from 11,800 to 4750 cal yr BP, with large paleofires between 7450 and 4750 cal yr BP and between 1300 and 70 cal yr BP. In Central Brazil, paleoclimatic changes occurred, with a reduction in precipitation between 7500 and 7000 ^{14}C yr BP (~8338 and 7697 cal yr BP) (Salgado-Labouriau et al., 1998; Barberi et al., 2000) and between 7500 and 5530 ^{14}C yr BP (~8338 and 6212 cal yr BP) (Behling, 1995a). Dry conditions during the Holocene by 3500–1400 ^{14}C yr BP (~3826–1268 cal yr BP) in the Chaco region were recognized by Iriondo and Garcia (1993). In the Paraná River Hydrographic Basin, a probable drier paleoclimatic phase between 4640 and 4360 ^{14}C yr BP (~5442 and 4832 cal yr BP) was described (Stevaux, 2000).

In Southern and Southeastern Brazil, where today there are *Araucária* forests, there was predominance of ground vegetation during the last glacial period (cold and relatively dry climatic conditions), and decreased and lowland forest taxa increased during the late-glacial period (wetter and warmer conditions) (Behling, 2002). The initial expansion of the *Araucária* forests from gallery forests occurred between 2850 and 1530 ^{14}C yr BP (~2958 and 1315 cal yr BP), indicating wetter climate conditions with a shorter dry season (Behling, 1997). In semi-deciduous forest regions, there were large areas of "cerrado" during the early Holocene, with an abundance of paleofires, indicating a drier climate (De Oliveira, 1992; Behling, 1995b; Parizzi et al., 1998; Rodrigues Filho et al., 2002). Areas occupied by the *Araucaria* forests in the last glacial maximum (LGM) were gradually replaced by "cerrado" and by mesophytic forests from 10,000 ^{14}C yr BP (11,404–11,249 cal yr BP), until the present day (De Oliveira, 1992; Ledru et al., 1996). In Southern Brazil, wetter climatic conditions were observed from 5170 cal yr BP, with an even wetter period from 1550 cal yr BP (Behling et al., 2005). Extensive peat bog formation began between 9720 and 8853 ^{14}C yr BP (~11,185 and 9695 cal yr BP), indicating a wetter and colder climate between 9700 and 8200 ^{14}C yr BP (~11,175 and 9900 cal yr BP) than at present. It was also hotter than at present between 8200 and 5400 ^{14}C yr BP (~9900 and 6000 cal yr BP), and afterwards, it was drier (Garcia et al., 2004).

In the coastal area of Southeastern Brazil, there is evidence that in 8000 ^{14}C yr BP (8978–8646 cal yr BP), the climate became hotter and wetter (Luz et al., 1999; Coelho et al., 2002; Coelho et al., 2008). In approximately 7000 ^{14}C yr BP (7835–7697 cal yr BP), fluctuations in the sea level formed numerous lagoons in the coastal region. Wetter conditions can be identified between 6300 and 4650 ^{14}C yr BP (~7254 and 5286 cal yr BP), and some data point to drier conditions in 6000 ^{14}C yr BP (6847–6713 cal yr BP). With the lowering of the sea level in 5100 ^{14}C yr BP (5894–5730 cal yr BP), a semi-deciduous tropical forest and an ombrophilous forest developed. There was a drier phase with higher temperatures between 4650 and 1350 ^{14}C yr BP (~5286 and 1180 cal yr BP), followed by a wetter phase after 1350 ^{14}C yr BP (1283–1180 cal yr BP), with a drier interval between 775 and 213 ^{14}C yr BP (~678 and 167 cal yr BP) (Luz et al., 1999; Coelho et al., 2002; Coelho et al., 2008). In Southern Brazil, the marine influence on coastal lagoons was determined to have occurred in 9620 and 7370 ^{14}C yr BP (~11,083 and 8038 cal yr BP) in a relatively dry climate, with initial transgression between 7185 and 5575 ^{14}C yr BP (~8005 and 6282 cal yr BP), maximum transgression from 5575 to 4480 ^{14}C yr BP (~6282 to 4875 cal yr BP), followed by regression between 4480 and 3325 ^{14}C yr BP (~4875 and 3444 cal yr BP) (Dillenburg and Hesp, 2009).

Anthracological studies indicate that coastal ecosystems are strongly conditioned by edaphic factors and are minimally influenced by climatic fluctuations (Scheel-Ybert, 2000). Through several proxies, variation curves of paleosea levels over the last 7000 ^{14}C yr BP (7835–7697 cal yr BP), were interpreted for the Brazilian coast (Martin et al., 1996; Tomazelli and Villwock, 1996). The Salvador curve (Bahia) shows more than 50 changes of the paleo sea level, and it indicates that the maximum sea level occurred about 5000 ^{14}C yr BP (5721–5607 cal yr BP). It also shows two negative oscillation levels when the paleo sea level was below the current level between 4000 and 3800 ^{14}C yr BP (~4442 and 3990 cal yr BP) and between 3000 and 2700 ^{14}C yr BP (~3208 and 2743 cal yr BP), with dried estuaries and lagoons replaced by freshwater swampy lowlands (Martin et al., 1996).

Based on the dating of fossil vermetid tubes collected above the current sea level, a new curve for the last 5500 ^{14}C yr BP (6293–6208 cal yr BP) was outlined without negative oscillations for the Holocene paleo sea levels (Angulo and Lessa, 1997; Angulo et al., 1999). This created much debate regarding the amplitude of sea level oscillations and the correct interpretation of data and the construction of curves in Brazil (Lessa and Angulo, 1998; Martin et al., 1998; Angulo et al., 2006; Mahiques et al., 2010).

The Salvador curve was reconstructed by Martin et al. (2003) with new evidences, corrections for isotopic fractionation and reservoir effects and calibrations for astronomical ages. At about 7800 cal yr BP, the mean sea level exceeded the current level for the first time in the Holocene. The mean sea level exceeded the present level by 4.7 ± 0.5 m by 5600 cal yr BP, followed by a fast regression between 5300 and 4200 cal yr BP, when the mean sea level may have been below the current level. Fast regression occurred again about 3700 cal yr BP, with a maximum of

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3.5 ± 0.5 m above the actual mean sea level, followed by a regular and slow decrease between 3500 and 2800 cal yr BP. About 2800 cal yr BP, a very fast decline of mean sea level occurred, falling to below the current level by 2600 cal yr BP. About 2300 cal yr BP, the mean sea level began to rise, reaching 2.3 ± 0.5 m above the present level by 2100 cal yr BP. Since then, mean sea level dropped regularly to its present position (Martin et al., 2003).

A sea level higher than present (± 5 m) at the Rio de Janeiro coast did not correspond to the presence of shell mounds 7950–7860 ^{14}C yr BP (8776–8543 cal yr BP) (Anjos et al., 2010). Although the sea level provided wetlands, which created a favorable environment for the occupation and construction of shell mounds, the area was not submerged at that time. Available stratigraphic data for the São Paulo coast indicate that sea levels were below the current sea level as much as −13 m about 8000 ± 50 cal yr BP and about −6 m by 7470 ± 60 cal yr BP (Mahiques and Souza, 1999; Klein, 2005; Mahiques et al., 2010). Archeological data gathered by Ybert et al. (2003) indicate that between 4900 and 3470 ^{14}C yr BP (~5615 and 3607 cal yr BP), there was a lagoon on the São Paulo coast surrounded by a relatively open forest. This indicates that the sea level at that time was higher than at present and contradicts the negative oscillation data from Martin et al. (1996). When the sea level fell about 3470 cal yr BP, the lagoon was replaced by a swamp forest that still exists today, with a short, wetter period between 1300 and 675 cal yr BP (Ybert et al., 2003). By 1300 ^{14}C yr BP (1263–1167 cal yr BP), a swamp already existed on the coast, which expanded in 1000 ^{14}C yr BP (917–806 cal yr BP) and reached its present size in 700 ^{14}C yr BP (655–567 cal yr BP) (Amaral et al., 2006).

Palynologic studies show a strong correlation with the data from Martin et al. (1996), with the three periods of low pollen concentration connected to lowering of sea level: ~4200 ^{14}C yr BP (~4821–4577 cal yr BP), ~2200 ^{14}C yr BP (~2299–2052 cal yr BP) and ~700–200 ^{14}C yr BP (~655–147 cal yr BP) (Barth et al., in press). Data from a Southeastern Brazilian platform indicate environmental changes associated with the early transport of cold and less saline waters towards the North (Plata Plume Water) from 5200 to 5000 cal yr BP and from 3000 to 2800 cal yr BP and an increase in atmospheric moisture by 1500 cal yr BP (Mahiques et al., 2009).

Study area

The Juréia-Itatins Ecological Station, one of the main groups of protected South Atlantic ecosystems, extends over an area of 600 km^2, to the North of the Cananéia-Iguape lowlands in the Southeast side of the state of São Paulo. During the Neogene, paleoclimatic variations had an important role in the region's morphogenesis and sedimentation, and the evolutionary history. Phases distributed between the high Pleistocene and the end of the Holocene, included lowered areas that show marine, river, lagoon and colluvial deposits (Suguio and Martin, 1978; Villwock et al., 1986; Tomazelli and Villwock, 1996; Tomazelli et al., 2000).

In this conservation unit, Precambrian crystalline basements, Pleistocene beach sedimentary deposits (Cananéia Formation) and Holocene (Ilha Comprida Formation) deposits of paleolagoon origin can be found. The paleolagoon deposits are separated from the open sea by a strip of 3 to 5 km of sandy beach ridges (beach crests).

At the height of the LGM, a transgression (Santos transgression) occurred about 17,500 ^{14}C yr BP (21,227–20,488 cal yr BP), and formed a bay at Juréia that provided for the proliferation of mollusks (Suguio and Martin, 1978; Martin et al., 1996). With the formation of a barrier island about 5150 ^{14}C yr BP (5912–5750 cal yr BP), a broad lagoon formed. The lagoon gradually disconnected with the open ocean, but developed a large lagoon system with increasingly salty water. Currently, the area of paleolagoon sediments is drained by rivers that flow into the Atlantic Ocean, and several shell mounds, linked to the Holocene evolution of the paleolagoon, are found at its banks.

Methods

Paleolagoon sediments were subjected to multi-proxy analyses of textural, mineralogical, geochemical and isotopic compositions at a regular sampling interval. The 5.79 m core, identified as S03 was collected using vibracore equipment (24°29′16.45″S/47°15′41.09″W, Figs. 1 and S1, Supplementary movie). Samples were collected at 10 cm intervals and subsamples were also collected at 2 cm intervals in the first meter because of the difference in sedimentation rates observed throughout the core sample.

A high-precision georeferenced topographic survey was performed in relation to the current relative sea level at the coring site (see Supplementary). Core S03 contains 5.79 m of massive paleolagoon sediments, the top of which is situated at an elevation of 2.63 m above sea level (present altitude) and 8 km from the coast. Data and deposition times were interpolated to pseudo-annual intervals and then further converted by averaging over longer time intervals. The distribution and variability of grain size as well as certain elements and their isotopes, were used as proxies of environmental changes in the source area.

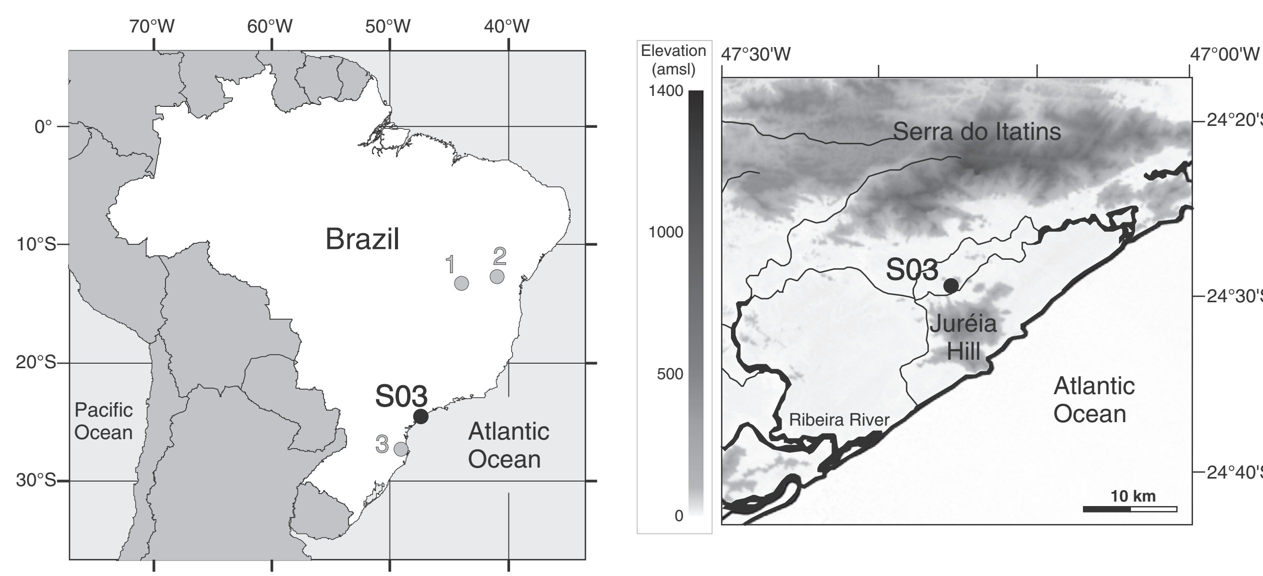

Figure 1. Map of the study area showing the location of core S03 (24°29′16.45″S/47°15′41.09″W — marked with black circle), the State of São Paulo, Brazil on Juréia-Itatins Ecological Station (amsl = above mean sea level). The speleothem records from Brazil are marked with a gray circle: Padre Cave (1) and Paixão Cave (2) (Cheng et al., 2009), Botuverá Cave (3) (Cruz et al., 2005).

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The chronology of the paleolagoon is based on ages of fifteen bulk sediment samples from core S03 determined by accelerated mass spectrometry (AMS) radiocarbon dating, performed by Beta Analytic Inc., USA. This study uses conventional time-stratigraphic nomenclature where radiocarbon (^{14}C) ages are given as years before present (^{14}C yr BP) referenced to AD 1950. Radiocarbon ages were converted to calibrated ^{14}C ages from INTCAL04 (Talma and Vogel (1993) and Reimer et al., 2004), which involves a two-sigma error in the radiocarbon measurements with an error multiplier of 1.0. In turn, the calibrated ages were recalculated using the Southern Hemisphere Calibration with SHCAL04 (McCormac et al., 2004). A geochronological model for the core was built based on the calibrated radiocarbon dates (median probability) with adjustments for depth and changes in sedimentation rates along the core.

Conventional grain size determination in sediments was performed at the University of São Paulo, Brazil (IGc-USP), along with identification of clay minerals by X-ray diffraction at the LAMIR-UFPR, Brazil, and heavy mineral content at the University of São Paulo, Brazil (LCT-USP). At the SGS GEOSOL Laboratory, measurements of the trace elements were made using Inductively Coupled Plasma Mass Spectrometry (ICP-MS), major elements were identified using X-ray fluorescence (XRF), and carbonate content was estimated by loss on ignition (LOI) at 1000°C.

For the determination of total carbon content (TC), total organic carbon (TOC), total nitrogen (TN), total organic nitrogen (TON), total sulfur (TS) and the ^{13}C, ^{15}N and ^{34}S stable isotopic composition of sediments, samples were combusted on an ECS 4010 elemental analyzer (Costech Analytical, Valencia, CA). The resultant CO_2, N_2, or SO_2 gasses were measured on a Delta Plus XP isotope ratio mass spectrometer (ThermoFinnigan, Bremen). These measurements were performed at the Stable Isotope Core Laboratory at Washington State University, USA.

Carbon isotope results are reported in per mil relative to Vienna Peedee Belemnite (VPDB) by assigning a value of +1.95 per mil to NBS 19 $CaCO_3$ and −46.6 per mil to LSVEC Li_2CO_3 (Coplen et al., 2006). The nitrogen isotope ratios are reported in parts per mil relative to N_2 in the air (Coplen, 1994). Final ^{13}C and ^{15}N results were calculated with a three-point correction consisting of running standards calibrated to a suite of international reference materials. The typical precision of the analyses was <0.2 per mil for ^{13}C and <0.4 per mil for ^{15}N (1SD). Sulfur isotope ratios are expressed in parts per mil relative to Vienna Cañon Diablo Troilite (VCDT) by assigning a value of −0.3‰ exactly to IAEA-S-1 (Krouse and Coplen, 1997). Sulfur analyses were performed with a two reactor configuration consisting of copper wire and quartz chips (Fry et al., 2002). Approximately 5 mg of Nb_2O_5 were added to every sample as a combustion aid. The final ^{34}S results were calculated using a five-point normalization consisting of IAEA-S-2, IAEA-S-3, IAEA-SO5 and two running standards calibrated to a suite of international reference materials. The typical precision of the analyses was <0.5 per mil for ^{34}S.

All elemental estimates of total C, N and S were calculated from a multi-point correction using acetanilide or sulfanilamide, respectively. Inorganic C was removed with three successive washes of ^3N phosphoric acid. With the goal of understanding relative changes in terrestrial versus aquatic matter in the sediments, C:N atomic ratios were obtained (Bengtsson and Enell, 1986; Meyers and Teranes, 2001).

The Pb isotopic compositions of the sediment samples were measured at the Center of Geochronological Research, University of São Paulo, Brazil (IGc-USP). About 50 mg of sediment were digested with an acid mixture of HF, HNO_3 and HCl for 5 days and Pb was purified using the standard ion exchange technique, following the procedure of Babinski et al. (1999). Pb isotope ratios were measured by solid-source thermal ionization mass spectrometry technique (TIMS) using a multicollector Finnigan MAT 262 mass spectrometer. Measured ratios of the Common Pb NBS-981 standard were in close agreement with their certified values.

Results and discussion

Core S03 is composed of massive dark mud paleolagoon sediments (Fig. S2), themselves composed of three plastic silty clay units with vegetable fragments at the following depths: top down to 0.24 m (brown), 0.24 to 0.42 m (dark yellow) and 0.42 m to the base (dark gray). No structures or marks were found in the rest of the core.

Table 1
AMS (Accelerator Mass Spectrometry) radiocarbon ages from the paleolagoon sediments of core S03. Analyses were performed by Beta Analytic, Inc.

Depth (m)	Lab. number	$^{13}C/^{12}C$ (‰)	Conventional ^{14}C age BP	Calendar age cal yr BP (median probability)[a]	Calendar age range (cal yr BP)[b]	Elevation a.m.s.l.(m) [c]
0.00	Beta-238846	−23.9	108.1 ± 0.4 pMC[d]	Modern	Modern	+2.63
0.10–0.12	Beta-245792	−23.1	280 ± 40	305	460 to 280	+2.53
0.30–0.32	Beta-245793	−25.4	2180 ± 40	2120	2090 to 2060 2320 to 2100	+2.32
0.38–0.40	Beta-245794	−25.0	3890 ± 40	4240	4200 to 4160 4420 to 4230	+2.25
0.50–0.52	Beta-270086	−25.6	4190 ± 40	4640	4840 to 4780 4770 to 4580	+2.13
0.60–0.62	Beta-245795	−24.2	5440 ± 40	6205	6300 to 6190	+2.03
0.80–0.82	Beta-245796	−24.4	6620 ± 50	7450	7580 to 7430	+1.83
0.90–0.92	Beta-245797	−24.0	6760 ± 50	7575	7680 to 7560	+1.73
1.00–1.02	Beta-244060	−24.2	6830 ± 50	7615	7750 to 7580	+1.63
1.70–1.72	Beta-270087	−23.6	7220 ± 50	7975	7950 to 8170	+0.93
2.00–2.02	Beta-244061	−24.1	7600 ± 50	8375	8460 to 8350	+0.63
3.00–3.02	Beta-244062	−24.1	7620 ± 50	8385	8520 to 8360	−0.37
4.00–4.02	Beta-244063	−24.1	7810 ± 50	8545	8700 to 8670 8650 to 8510 8500 to 8470	−1.37
5.00–5.02	Beta-245798	−23.7	7970 ± 50	8700	9010 to 8630	−2.37
5.79	Beta-238847	−24.2	8370 ± 50	9400	9490 to 9280	−3.16

BP = before present, AD 1950; cal = calibrated.
[a] Calibrated ages are calculated from SHCAL04 (McCormac et al., 2004).
[b] Calibrated ages are calculated from INTCAL 04 (Reimer et al., 2004) and Talma and Vogel (1993), which assumes a two-sigma error on radiocarbon measurements with an error multiplier of 1.0.
[c] a.m.s.l. = actual altitude above mean sea level.
[d] pMC = reported data indicates an age of post BP and has been assumed as a% of the modern reference standard, which suggests that the material came from a living being within the last 50 yr.

Fifteen AMS radiocarbon dates were determined with the bulk sediment of core S03 (Table 1). The data and deposition times were linearly interpolated to pseudo-annual intervals, and then, these data were converted by averaging over longer intervals of time. Based on the age model, the basal age of the core is 9400 cal yr BP, and the core top is assigned to the modern age (Fig. 2).

The sedimentation rate obtained over the 9400 cal yr BP showed high values with a significant degree of variability (Fig. 3): 0.004 to 0.016 cm/yr (0 to 0.8 m), 0.08 to 0.25 cm/yr (0.80 to 2 m), 10 cm/yr (2 to 3 m), 0.625 to 0.645 cm/yr (3 to 5 m) and 0.11 cm/yr (5 to 5.79 m). The high rate of sedimentation between 9400 (5.79 m) and 7450 cal yr BP (0.80 m) made it possible to obtain a high-resolution sedimentary record of events for this period. Very high anomalous values were observed for the period between 8385 and 8375 cal yr BP (3–2 m).

Kaolinite and illite are present in all the analyzed samples (Supplementary Table S1). Montmorilonite is only absent between 0.0 and 0.22 m, when a sedimentary change occurs (Fig. S2), and between 1.60 and 1.72 m. Pyrophyllite is observed only at 5.40 m (9054 cal yr BP) and 2.10 m (8376 cal yr BP). Hydrobiotite and vermiculite occur between 0.00 and 0.20 m. Most of the clay mineralogy occurrences do not indicate any change in the environment from the full record, with the constant presence of clay minerals related to pedogenetic processes and commonly found in continental and transitional environments.

Clay, silt and sand, as well as heavy mineral content (Supplementary Tables S2 to S4), were not constant from present day to 9400 cal yr BP, indicating important changes in sediment inflow. We observed a marked change in grain size between 7615 and 8545 cal yr BP, reflected in peaks of parameters such as the average diameter, the standard deviation, asymmetry and kurtosis.

At approximately 8300 cal yr BP, low values occur for sand, silt and heavy minerals in fine sand. At the same time, the fine:coarse ratio and content of heavy minerals in very fine sand increase, suggesting a lower energy in the environment and a decrease of the continental contribution to sedimentation. At approximately 6300 cal yr BP, the silt and clay content decreased, diminishing the fine:coarse ratio and indicating a higher energy for the environment.

Variability in the ratios between chemical elements indicates mineral diversification and availability in the source area and suggests environmental changes (Supplementary Tables S5 and S6, Figs. 5 and 6). Elements such as Ag, Bi, Cd, Sb, Se, Sn, Th, Tl and U were not detected.

Around 9400 cal yr BP, some elements showed decreasing concentrations throughout the sedimentary record (Al_2O_3, TiO_2, Fe_2O_3, MgO, As, Co, Cu, Y, Pb, Sc, La and Fe), indicating a lower degree of input of terrestrial material into the paleolagoon, whereas others showed the reverse behavior (K_2O, Na_2O, SiO_2, Na), which indicates a greater input of marine elements to the paleolagoon. Despite showing low values, the changes in sodium content indicate a saline wedge influence caused by brackish water invasion.

Between 9400 and 8500 cal yr BP, Cu, LOI, TOC, TON, TC and TN values continuously decrease, which reflects a reduction in organic compounds from biological activity. However, at about 8500 cal yr BP, there are abrupt increases in TOC, TON, TC and TN values that reflect major increases in organic productivity.

Values of SiO_2/Al_2O_3 show a pronounced increase about 9400 cal yr BP, with a subsequent decrease about 8400 cal yr BP, indicating marine contribution in this period. The opposite behavior is found for the Co, Fe and Fe_2O_3 concentrations, which suggests a greater degree of variation in the pH and Eh (redox) conditions in the paleolagoon due to contributions of marine and continental origins along the record.

Between 8550 and 8480 cal yr BP (3.60–4.00 m), the lowest and/or highest concentrations of some elements in the whole record

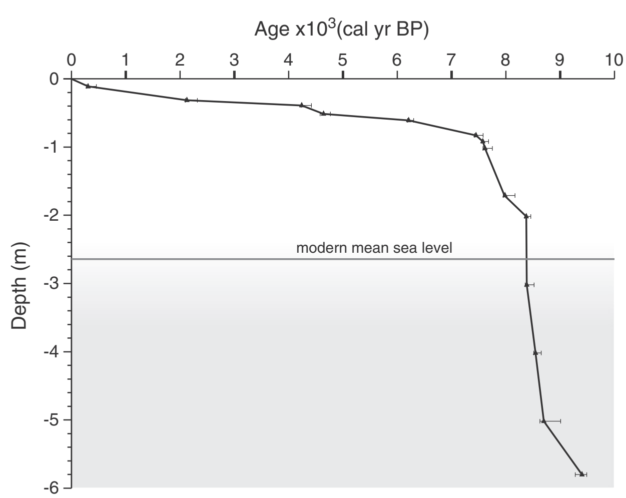

Figure 2. Ages versus depth, core S03. Age model was developed using calibrated radiocarbon ages, calculated from SHCAL04 (McCormac et al., 2004), which are shown with two sigma error bars, calculated from INTCAL 04 (Reimer et al., 2004).

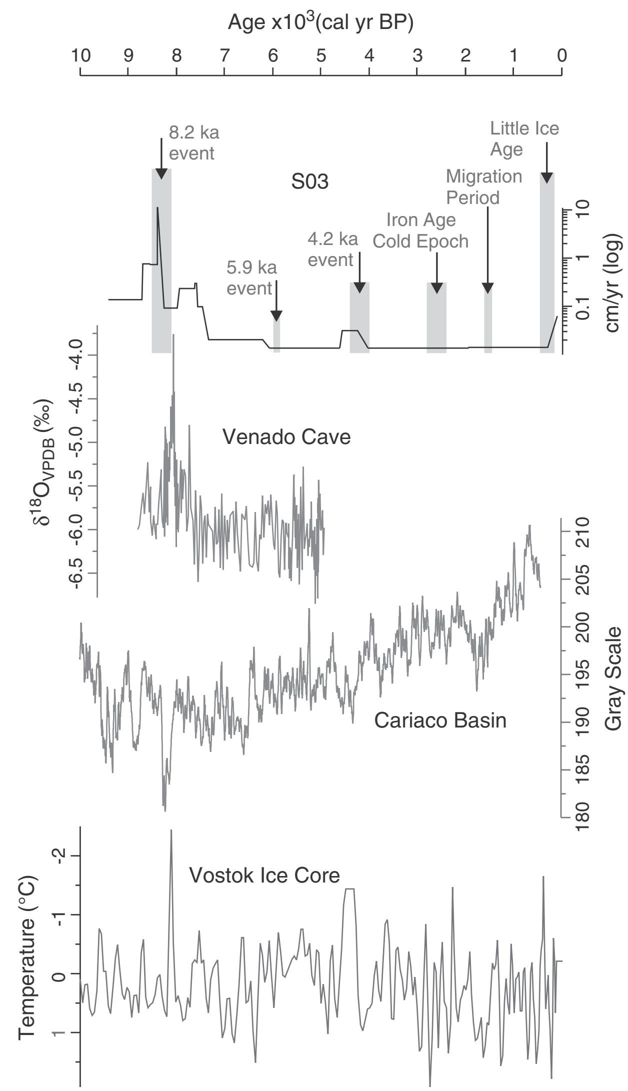

Figure 3. The rate of sedimentation (cm/yr) of core S03 shows an anomaly: very high values were observed for the period between 8385 and 8375 cal yr BP (3 to 2 m), that is contemporaneous with records elsewhere of the 8.2 ka event (Bond et al., 1997, 2001). The speleothem stable isotope data ($\delta^{18}O_{VPDB}$) are from Venado Cave, Costa Rica (Lachniet et al., 2004). The sediment grayscale record is from the Cariaco Basin, Venezuela (Hughen et al., 2000). The temperature difference with respect to the mean recent time value is from the Vostok Ice Core, Antarctica (Petit et al., 1999).

are observed, which is followed by a sudden increase/decrease in the concentrations that lasts for at least 500 years. Around 8500–8400 cal yr BP (3.80–3.70 m), substances such as CaO, Ba, Ca, Sr, P_2O_5, SiO_2, P and W show significant increases in their concentrations, which suggest a higher degree of marine influence on the paleolagoon (Supplementary Table S5, Fig. 4).

Other substances (Cr, Li, Ni, Zr, Al, V, Sc, Zn, K, Mg, Mo, V, Sc, Ti, Fe_2O_3 and MgO) show increased concentrations between 8400 and 7575 cal yr BP, which suggests a higher level of terrestrial detritic input, also suggested by values of the C:N ratio. A more uniform contribution of mafic minerals in the source area between 8500 and 7600 cal yr BP suggests a higher continental contribution to sediments.

We observed very low values of arsenic and scandium between the present day and 2544 cal yr BP and between 8700 and 8529 cal yr BP. Low magnesium volumes between the present day and 5200 cal yr BP were also observed (Supplementary Table S6, Fig. 5).

Finally, low volumes of calcium, cobalt, sodium and lanthanum between the present day and 4300 cal yr BP were observed. The decreases in Zr/Cr values since 4200 cal yr BP and Zr values since 2500 cal yr BP suggest a lower continental contribution.

The TOC, TON and TS contents of the sediments were low throughout the record, but the variation between maximum and minimum contents indicates paleoenvironmental changes over the entire period since 9400 cal yr BP. The TOC content was lower than 3% for most of the sediments, with values between 0.65 and 5.99%. The TON content was lower than 0.15% for most of the sediments, with values between 0.04 and 0.32%, which shows a correlation with the TOC behavior. The TS content was lower than 2% for most of the sediments, with values between 0.015 and 2.868%.

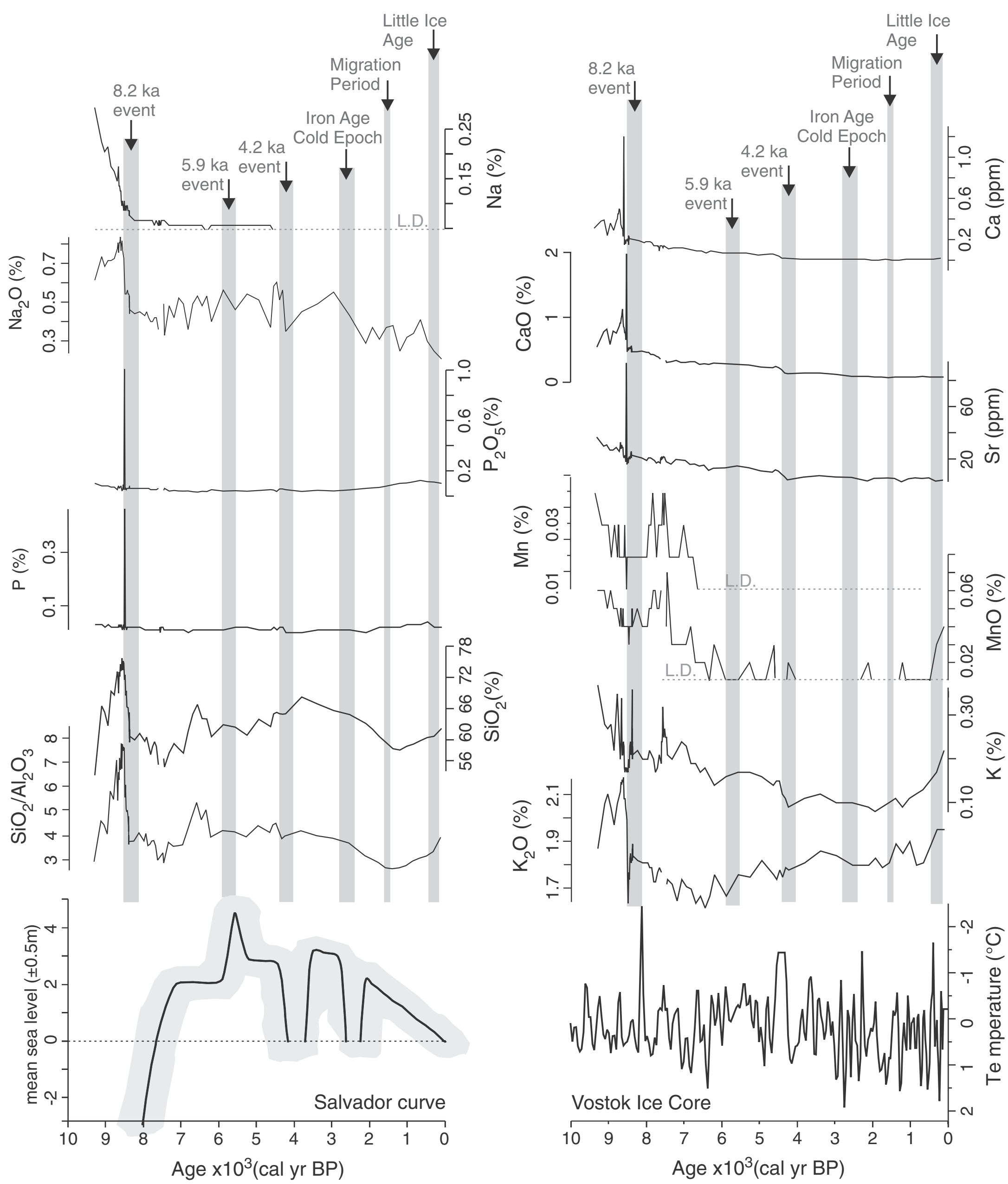

Figure 4. The sequence of geochemical proxy records of core S03 shows anomalies contemporaneous with Bond events (Bond et al., 1997, 2001). The sea-level data are from the Salvador curve (Martin et al., 2003). The temperature difference with respect to the mean recent time value is from the Vostok Ice Core, Antarctica (Petit et al., 1999).

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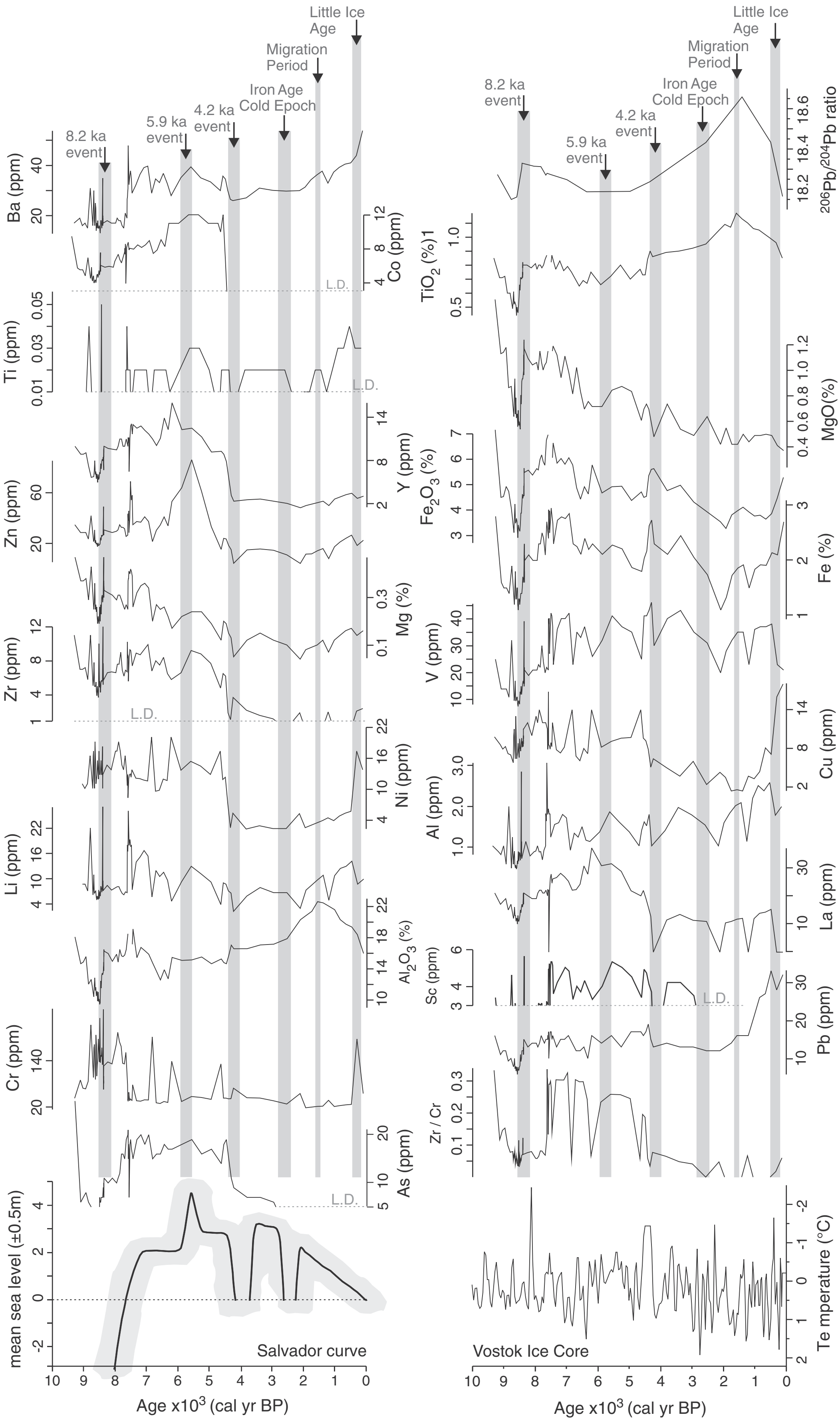

Figure 5. The geochemical sequence of proxy records of core S03 shows anomalies contemporaneous with Bond events (Bond et al., 1997, 2001). The sea-level data are from the Salvador curve (Martin et al., 2003). The temperature difference with respect to the mean recent time value is from the Vostok Ice Core, Antarctica (Petit et al., 1999).

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Well-defined peaks for TOC, TON and TS suggest that abrupt paleoenvironmental changes occurred (1) 8500–8300 cal yr BP; (2) 8000–7800 cal yr BP; (3) 4600–4200 cal yr BP; (4) 1400–1300 cal yr BP; and (5) 300 cal yr BP (Tables 2 and 3, Supplementary Tables S7 and S8, Fig. 6), probably correlated to Bond events (Bond et al., 1997, 2001).

The C:N ratios are between 10 and 20 for most of the sediments, ranging from 6.05 (4320 cal yr BP) to 39.93 (7769 cal yr BP) throughout the core, indicating that organic matter from phytoplankton and C3 terrestrial origin plants accumulated, but with episodes of greater contribution from marine phytoplankton and organic matter of mixed origin.

The C:N ratios indicate the presence of organic matter of algal origin from 4320 cal yr BP to the present day and about 6952 cal yr BP, indicating a higher marine contribution during those times. Sediment changes on top of the core marked 4320 cal yr BP (0.42 m depth, Fig. S2). Organic matter of mixed origin can be found between 4320 and 6952 cal yr BP, between 7076 and 7717 cal yr BP and between 8377 and 9338 cal yr BP, indicating marine and continental contributions to the lagoon environment, with high contributions of marine elements around 8385 cal yr BP.

Between 7769 and 8376 cal yr BP, the C:N ratios indicate the presence of C3 vascular vegetation, suggesting deposition of organic carbon. An increase in the TOC content suggests a major input of terrestrial material to the paleolagoon.

Throughout the core, $\delta^{13}C_{VPDB}$ values ranging from −24.05‰ to −26.84‰ indicate the presence of organic matter derived from C3 terrestrial plants and algae; $\delta^{15}N_{AIR}$ values of +4.68 to +8.85‰, between +5 and +6 for most of the sediments, indicate the presence of organic matter derived from phytoplankton; and $\delta^{34}S_{VCDT}$ values of −18.94 to +4.3‰ indicate the presence of organic matter derived from land plants. The organic matter is characteristic of continental origin with great influence of fresh water in the coastal environment (Fig. S3).

The interaction of $\delta^{15}N_{AIR}$ and $\delta^{13}C_{VPDB}$ demonstrates that the organic matter of paleolagoon sediments is predominantly derived from a mixture of sources, which includes C3 terrestrial plants and phytoplankton. The interaction of $\delta^{34}S_{VCDT}$ and $\delta^{13}C_{VPDB}$ demonstrates that the organic matter of paleolagoon sediments is predominantly derived from terrestrial plants.

All isotopes of C ($\delta^{13}C_{VPDB}$), N ($\delta^{15}N_{AIR}$) and S ($\delta^{34}S_{VCDT}$), as well as TOC, TON and TS show the following well-defined peaks: (1) 8500–8300 cal yr BP; (2) 8000–7800 cal yr BP and (3) 4600–4200 cal yr BP (Tables 2 and 3; Supplementary Tables S7 and S8).

The values of $\delta^{15}N$ indicate strong marine contribution between 2120 and 1575 cal yr BP and a continental contribution about 7700 cal yr BP. The isotopic enrichment between 4240 and 1031 cal yr BP indicates a decrease in the contribution of organic matter derived from terrestrial plants. A slight enrichment of $\delta^{13}C_{VPDB}$ between 8490 and 8400 cal yr BP, between 9000 and 8520 cal yr BP and between 7525 and 7450 cal yr BP suggests a decrease in the contribution of organic matter derived from terrestrial plants.

The lead isotope ratios ($^{206}Pb/^{204}Pb$, $^{206}Pb/^{207}Pb$, $^{208}Pb/^{206}Pb$) (Table 4) vary throughout the whole record, but with persistent high ratios of $^{206}Pb/^{204}Pb$ occurring between 8500 and 7775 cal yr BP (Fig. 7). This pattern of change separates two distinct periods that characterize lead isotope ratios: (1) between 9161 and 8548 cal yr BP, when the $^{206}Pb/^{204}Pb$ ratio decreased from 18.273 to 18.161 and lead concentrations varied between 7.8 and 6.9 μg/g; and (2) between 8388 and 7780 cal yr BP, when the $^{206}Pb/^{204}Pb$ ratio increased to 18.305 by 8388 cal yr BP and varied only slightly (average = 18.315 ± 0.010). The same behavior is observed for lead concentrations (average = 14.0 ± 1.4 μg/g), which suggests that the

Table 2
The carbon isotopic compositions and content of the paleolagoon sediments of core S03.

Depth (m)	Calendar age (cal yr BP)	$\delta^{13}C_{VPDB}\times 1000$[a]	TOC (%)[b]	Depth (m)	Calendar age (cal yr BP)	$\delta^{13}C_{VPDB}\times 1000$[a]	TOC (%)[b]
0.00–0.02	108	−25.1	3.40	2.80–2.82	8383	−25.1	2.29
0.10–0.12	305	−24.9	2.17	2.90–2.92	8384	−25.0	2.46
0.20–0.22	1212	−25.4	2.22	3.00–3.02	8385	−25.0	3.60
0.30–0.32	2120	−26.8	1.00	3.10–3.12	8401	−24.5	1.53
0.40–0.42	4240	−25.6	0.82	3.20–3.22	8417	−24.5	1.55
0.50–0.52	4640	−26.3	2.75	3.30–3.32	8433	−24.4	1.52
0.60–0.62	6205	−26.4	2.97	3.40–3.42	8449	−24.4	1.46
0.70–0.72	6828	−25.4	2.58	3.50–3.52	8465	−24.9	1.48
0.80–0.82	7450	−24.9	2.49	3.60–3.62	8481	−24.8	1.14
0.90–0.92	7575	−25.0	2.71	3.70–3.72	8497	−25.8	1.75
1.00–1.02	7615	−24.6	2.54	3.80–3.82	8513	−25.2	1.30
1.10–1.12	7666	−25.3	2.69	3.90–3.92	8529	−24.8	0.76
1.20–1.22	7718	−25.3	3.06	4.00–4.02	8545	−24.5	0.95
1.30–1.32	7769	−26.0	5.99	4.10–4.12	8561	−24.7	0.72
1.32–1.34	7780	−25.2	5.03	4.20–4.22	8576	−24.9	0.86
1.40–1.42	7821	−25.1	3.98	4.30–4.32	8592	−24.5	0.73
1.50–1.52	7872	−25.2	2.84	4.40–4.42	8607	−24.4	0.87
1.60–1.62	7924	−25.5	3.33	4.50–4.52	8623	−24.4	0.91
1.70–1.72	7975	−25.1	3.04	4.60–4.62	8638	−24.4	0.79
1.80–1.82	8108	−25.3	3.11	4.70–4.72	8654	−24.2	0.89
1.90–1.92	8242	−25.2	3.21	4.80–4.82	8669	−24.2	1.03
2.00–2.02	8375	−25.4	3.15	4.90–4.92	8685	−24.1	1.01
2.10–2.12	8376	−25.2	3.24	5.00–5.02	8700	−24.8	0.72
2.20–2.22	8377	−24.7	2.43	5.10–5.12	8789	−24.3	0.75
2.30–2.32	8378	−25.3	2.75	5.20–5.22	8877	−24.7	1.46
2.40–2.42	8379	−25.5	2.43	5.30–5.32	8966	−24.4	1.75
2.50–2.52	8380	−25.1	2.24	5.40–5.42	9054	−24.8	1.55
2.60–2.62	8381	−24.8	1.93	5.50–5.52	9143	−24.8	1.55
2.70–2.72	8382	−25.1	2.19	5.72–5.74	9338	−25.2	2.74

[a] The carbon isotope results were reported in parts per mil relative to Vienna Peedee Belemnite (VPDB).
[b] TOC = total organic carbon; TC = total carbon.

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Table 3
The sulfur isotopic compositions and content of the paleolagoon sediments of core S03.

Depth (m)	Calendar age (cal yr BP)	$\delta^{34}S_{VCDT}$(‰)[a]	Sulfur (%)	Depth (m)	Calendar age (cal yr BP)	$\delta^{34}S_{VCDT}$(‰)[a]	Sulfur (%)
0.00–0.02	108	−0.29	0.039	2.80–2.82	8383	−10.08	1.69
0.10–0.12	305	−9.49	1.14	2.90–2.92	8384	−9.04	1.32
0.20–0.22	1213	4.30	0.031	3.00–3.02	8385	−1.99	0.052
0.30–0.32	2120	2.35	0.018	3.10–3.12	8401	−7.08	1.08
0.40–0.42	4240	−0.87	0.025	3.20–3.22	8417	−6.62	1.36
0.50–0.52	4640	−15.74	1.50	3.30–3.32	8433	−4.97	1.18
0.60–0.62	6205	−14.36	1.69	3.40–3.42	8449	−5.01	0.929
0.70–0.72	6828	−12.01	2.355	3.50–3.52	8465	−6.42	0.743
0.80–0.82	7450	−10.82	1.936	3.60–3.62	8481	−6.00	0.742
0.90–0.92	7575	−8.68	1.95	3.70–3.72	8497	−10.76	0.981
1.00–1.02	7615	−11.27	1.889	3.80–3.82	8513	−11.17	0.666
1.10–1.12	7666	−16.57	2.378	3.90–3.92	8529	−4.03	0.857
1.20–1.22	7718	−12.88	1.659	4.00–4.02	8545	−4.61	0.916
1.30–1.32	7769	−11.91	2.635	4.10–4.12	8561	−2.85	0.668
1.32–1.34	7780	−12.37	2.868	4.20–4.22	8576	−3.94	0.71
1.40–1.42	7821	−11.69	1.84	4.30–4.32	8592	−0.87	0.812
1.50–1.52	7872	−10.94	1.46	4.40–4.42	8607	−1.04	0.961
1.60–1.62	7924	−8.18	1.69	4.50–4.52	8623	−2.04	0.888
1.70–1.72	7975	−7.40	1.56	4.60–4.62	8638	−2.27	0.636
1.72–1.74	8002	−6.77	1.76	4.70–4.72	8654	−1.72	1.01
1.80–1.82	8108	−7.77	1.35	4.80–4.82	8669	1.28	0.964
1.90–1.92	8242	−8.01	1.68	4.90–4.92	8685	2.91	0.948
2.00–2.02	8375	−9.46	1.45	5.00–5.02	8700	1.07	0.655
2.10–2.12	8376	−10.61	2.05	5.10–5.12	8789	−0.72	0.721
2.20–2.22	8377	−7.26	1.39	5.20–5.22	8877	−1.38	1.07
2.30–2.32	8378	−8.18	1.51	5.30–5.32	8969	−3.91	1.16
2.40–2.42	8379	−9.96	1.14	5.40–5.42	9054	−2.87	0.986
2.50–2.52	8380	−9.66	1.16	5.50–5.52	9143	−4.04	1.05
2.60–2.62	8381	−11.00	1.14	5.58–5.59	9338	−4.99	0.885
2.70–2.72	8382	−9.44	1.51	5.72–5.74	8607	−18.94	1.89

[a] The sulfur isotopic ratios are reported in parts per mil relative to Vienna Cañon Diablo Troilite (VCDT).

sediment sources were significantly different between the two periods.

The 8.2 ka event

In the Hudson Bay region, on the eastern edge of North America, there were two very large glacial lakes which were held in place by a natural dam (Barber et al., 1999; Leverington et al., 2002). Approximately 8200 yr ago, the dam catastrophically failed and released an enormous amount of fresh water into the ocean. The seawater became less salty and lighter, which gave rise to a salinity anomaly in the North Atlantic that transformed the global thermohaline circulation. This occurrence had a significant impact on the climate, specifically the lowering of global temperatures in mainly the Northern Hemisphere (Barber et al., 1999). Within a relatively short period of time, climatic conditions changed; this period has come to be known as the 8.2 ka event (Kobashi et al., 2007; Thomas et al., 2007).

This event was recognized for the first time in Greenland ice cores (NGRIP, 2004), in which evidence was found for lower temperatures and accumulations of snow lasting more than two decades (Alley et al., 1997; Kobashi et al., 2007; Thomas et al., 2007). Air bubbles present in Greenland ice cores suggest a 10–15% drop in methane concentrations (Alley et al., 1997), which has been correlated with a decrease in the area covered by tropical rainforests brought about by a drier climate. There is evidence that the 8.2 ka event caused a temperature drop of around 2°C in Europe (Wick and Tinner, 1997), pronounced climate cooling from 8.9 to 8.3 ka in the USA (Hu et al., 1999), significant sea level changes in the Norwegian Sea (Klitgaard-Kristensen et al., 1998), extremely dry conditions on the southeastern edge of the Sahara (Gasse and Van Campo, 1994), glacial advances in the New Zealand Southern Alps (Salinger and McGlone, 1990), and an increase in precipitation and/or a decrease in temperature in South America and the central South Atlantic (Douglass et al., 2005; Ljung et al., 2008).

The fall in temperatures that accompanied the 8.2 ka event also corresponded with abrupt migrations of human populations and abandonment of sites ranging from Spain to Greece and in the Middle East (Gonzalez-Samperiz et al., 2009). Despite the fact that the 8.2 ka event is well-defined, its global extent is still not well-known because regional centennial-scale paleoclimatic records are still very scarce, especially in Brazil, where details are restricted to speleothem records (Cheng et al., 2009). Recent and new data presented herein indicate a paleoenvironmental anomaly in southeast Brazil that began 8500 cal yr BP and lasted 250–500 yr (Keigwin et al., 2005; Rohling and Pälike, 2005), which led to rapid changes in sediment deposition patterns on the Brazilian coast.

Due to the short duration of the effects related to the 8.2 ka event, approximately a few hundred years, the data obtained to the southeast of Brazil suggest that the event should have significantly affected environmental conditions in coastal. Such anomalies show a high degree of correlation with cooler conditions associated with the North Atlantic 8.2 ka event which appears to have affected Brazil by increasing local precipitation and humidity and sea level rise.

Ice cores from Greenland (Alley et al., 1997) and Africa (Thompson et al., 2002) suggest that the 8.2 ka event was global in extent. Additional data obtained to date indicate a correlation between paleoenvironmental changes in the South Atlantic and the 8.2 ka event: (1) proxy records of speleothems indicate that the South American Summer Monsoon underwent intensification in Brazil between 8300 and 8210 yr ago due to a change in the position of the Intertropical Convergence Zone (Cheng et al., 2009); (2) increased precipitation between 8275 and 8025 cal yr BP as a consequence of an increase in the sea surface temperature, which is indicated by sediment records from Nightingale Island in the central South Atlantic (Ljung et al., 2008); (3) glacial advance in the Cordilleras Real and Cochamamba (Bolivia) and Blanca (Peru) (Douglass et al., 2005; Glasser et al., 2009) as a result of an increase in sea-surface temperature in the South Atlantic and

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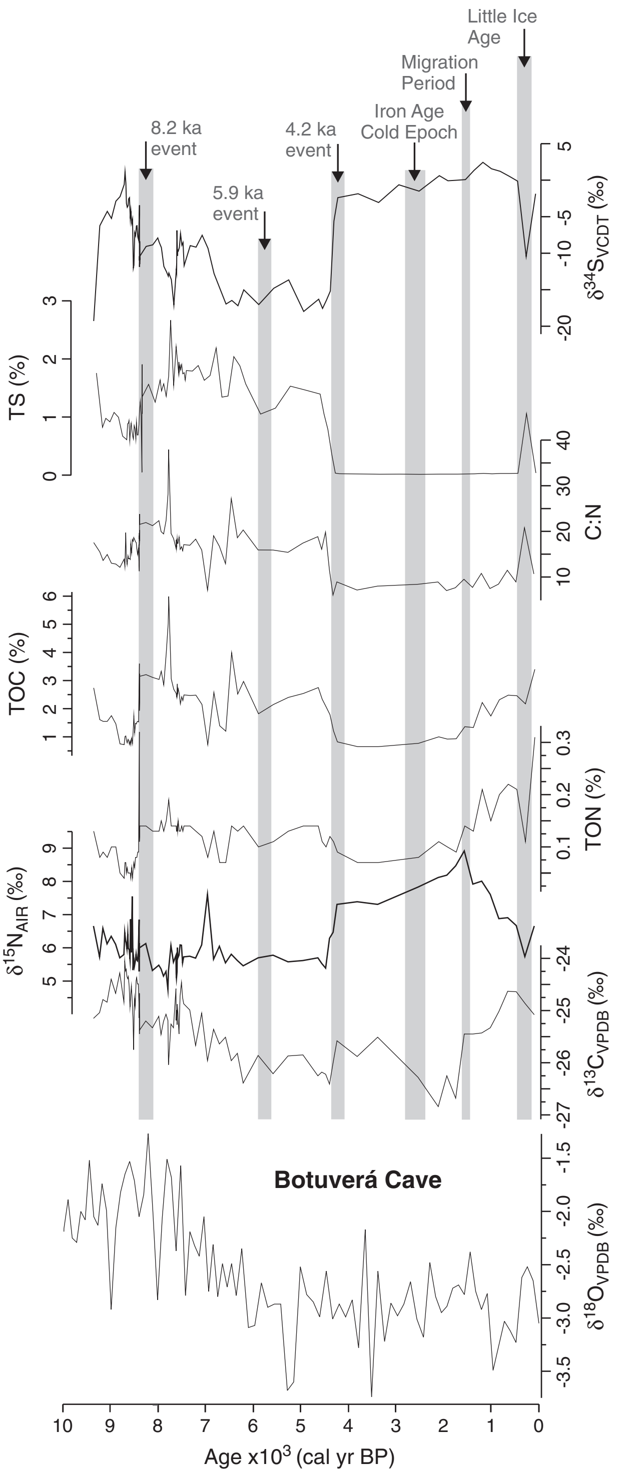

Figure 6. The values of total organic carbon (TOC), total organic nitrogen (TON) and total sulfur (TS), C:N ratios, pattern of stable isotopes of C ($\delta^{13}C_{VPDB}$), N ($\delta^{15}N_{AIR}$) and S ($\delta^{34}S_{VCDT}$) of core S03 show anomalies contemporaneous with Bond events (Bond et al., 1997, 2001). The speleothem stable isotope data ($\delta^{18}O_{VPDB}$) from Botuverá Cave, Brazil (Cruz et al., 2005).

intensification of the South American Summer Monsoon; (4) modeled sea level rise in Brazil (Kendall et al., 2008), and (5) the variability of the Juréia paleolagoon paleoenvironmental proxies reported in this study.

Other data for the same period of time but not correlated by the authors to the 8.2 ka event, substantiate the conclusion that the 8.2 event, although short and with a major environmental impact in Brazil, led to such paleoclimatic changes in the South Atlantic latitudes as: (1) drier phases in relation to the present in the Amazon area with widespread paleofires that began about 7600 cal yr BP (Cordeiro et al., 2008); (2) reduced precipitation in central Brazil 7500–7000 ^{14}C yr BP (~8338–7697 cal yr BP) (Salgado-Labouriau et al., 1998; Barberi et al., 2000); (3) significant positive excursion of $\delta^{18}O$ values about ~8196 yr ago with a higher value for the last 65 ka in Botuverá Cave, Brazil, indicated by speleothem proxy record (Cruz et al., 2005). Due to uncertainties in ^{14}C reservoir age corrections, small differences occur between the ages obtained in global records associated with the 8.2 ka event.

Conclusions

Several studies demonstrate that changes in atmospheric and oceanic circulation patterns, along with the climatic changes and glaciations that occurred in the Northern Hemisphere, were associated with environmental changes throughout the globe, including in South America. Those environmental changes are reflected in erosion, transport and sediment deposition processes observed over time, which are partially controlled by climatic patterns such as rainfall and changes in sea level. During short- and long-term climate transitions, fast reworking and deposition processes occur in sedimentary environments and create high-resolution records.

In this Brazilian study, we correlate anomalies in the sedimentary record with short- and long-term climate transitions and sea-level oscillations. Aggradation and degradation processes, which could be responsible for reworking sedimentary strata, were observed in the study area. However, episodes of paleoenvironmental change in the stratigraphic record appear to have been dominantly related to paleoclimatic transitions and sea-level changes.

Patterns of Holocene variability for paleoenvironmental proxy records representing sedimentation rate, grain size, mineral content, elemental abundance and isotopic variability indicate changes in dominant sedimentary origins over the last 9400 cal yr BP. Geochemical parameters and isotopic data (C, N, S and Pb) were very important for the identification of continental and marine influences in the paleoenvironment, with complex changes involving the intensity of paleoclimatic and sea level variation influences.

These changes in paleoenvironmental proxies are interpreted as reflecting a balance between a greater or lesser contributions from sediment sources within the Juréia paleolagoon drainage in response to changes in paleoclimatic conditions; these contributions involve on the one hand, factors such as humidity, precipitation and temperature, and on the other hand, an increase in sea level.

In Juréia about 9400 cal yr BP, a lagoon environment was established, with mixed continental and marine contributions. Short-term sedimentary episodes occurred about 8500 cal yr BP and are reflected by stratigraphic changes in through a high calcium concentration. Calcium had lowest content out of all the heavy minerals in the records, and its highest sedimentation rate occurred between 8385 and 8375 cal yr BP. Evidence suggests that the mean sea level reached or exceeded the current level in Juréia for the first time in the Holocene during that period.

By about 8375 cal yr BP, environmental change occurred, and the sedimentary record at Juréia shifted to dominance from continental sources. The change in dominance from continental sources is indicated by lead isotopes, which indicate stability in sea level, and also by increases in rainfall and sediment inflow. Between 8300 and 8210 cal yr BP, the South American Summer Monsoon underwent intensification in Brazil (Cheng et al., 2009). There is a very strong decrease in the sediment rate in Juréia between 8200 and 8000 cal yr BP, which is probably associated with higher rainfall and moisture. Additionally, dominance by

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Table 4
The Pb isotope data on sediments from the core S03.

Depth (m)	Calendar age (cal yr BP)	$^{206}Pb/^{204}Pb$[a]	Error (1σ)	$^{206}Pb/^{207}Pb$[a]	Error (1σ)	$^{208}Pb/^{206}Pb$[a]	Error (1σ)	Pb (ppm)
0.00–0.02	108	18.164	0.008	1.164	0.001	2.090	0.001	32
0.12–0.14	487	18.433	0.006	1.179	0.001	2.087	0.002	33
0.22–0.24	1394	18.658	0.007	1.192	0.001	2.082	0.002	16
0.32–0.34	2544	18.432	0.012	1.180	0.002	2.096	0.001	12
0.42–0.44	4320	18.237	0.005	1.167	0.001	2.114	0.001	15
0.52–0.54	4953	18.189	0.006	1.164	0.001	2.120	0.001	17
0.62–0.64	6330	18.187	0.007	1.164	0.001	2.121	0.003	10
0.72–0.74	6952	18.247	0.014	1.167	0.002	2.117	0.002	16
0.82–0.84	7475	18.273	0.006	1.169	0.001	2.115	0.001	15
0.92–0.94	7583	18.279	0.009	1.169	0.002	2.116	0.002	17
1.02–1.04	7625	18.270	0.011	1.170	0.003	2.116	0.004	15
1.32–1.34	7780	18.312	0.010	1.172	0.002	2.110	0.002	15
1.72–1.74	8002	18.314	0.007	1.172	0.002	2.113	0.002	14
2.02–2.04	8375	18.329	0.013	1.170	0.002	2.114	0.002	15
3.02–3.04	8388	18.305	0.011	1.171	0.003	2.110	0.003	12
3.48–3.50	8462	18.326	0.010	1.173	0.003	2.105	0.004	16
3.62–3.64	8484	18.351	0.005	1.174	0.001	2.105	0.001	16
3.88–3.90	8526	18.225	0.012	1.166	0.003	2.110	0.003	14
4.02–4.04	8548	18.161	0.008	1.164	0.002	2.115	0.002	6.9
5.02–5.04	8718	18.149	0.011	1.162	0.002	2.116	0.002	7.8
5.52–5.54	9161	18.273	0.009	1.168	0.002	2.114	0.002	11

[a] The Pb isotopic compositions are corrected for mass fractionation. The Pb blanks are lower than 100 pg.

continental organic matter until 7779 cal yr BP was identified by Ljung et al. (2008). The sedimentation rates clearly indicate more intensive phases and a better preservation of deposits around 8380 cal yr BP. This correlates with the occurrence of isotope and element variability, which indicates the existence of a greater intensity of sedimentary processes that may be correlated with the intense 8.2 ka event described in the literature. Global cooling off during the 8.2 ka event was temporary, but caused immediate climatic changes. Environmental influence of the increase in sea level and the resulting climatic changes continued in Brazil the event.

Between approximately 7717 to 7615 cal yr BP, the sediment rate dramatically decreased, and the contribution of marine elements to coastal transitional environment began. This change is probably due to more erosion, corroborating the data that indicate a lower sea level than the present between 8000 and 7470 cal yr BP (Mahiques and Souza, 1999; Klein, 2005; Mahiques et al., 2010) and the subsequent rise in sea level (Martin et al., 2003). However, the mean sea level did not exceed the current level at Juréia, and our data suggest that the rise in sea level brought only the creation of wetlands, thus corroborating data from Anjos et al. (2010). An environment dominated by marine elements occurs between 7076 and 6952 cal yr BP. By 6952 cal yr BP, a new lagoon environment with sedimentary contributions from marine and continental elements was occurring at Juréia, indicating a lowering of sea level, but not necessarily lower than the current level.

Between 5892 and 4640 cal yr BP, the sedimentation rate decreased significantly, and by 5579 cal yr BP there was a high content of elements that suggests a low level of terrestrial detritus (zinc and barium among others). The beginning of this period possibly is correlated with the 5.9 ka event, a climate cycle identified in the North Atlantic (Bond et al., 1997, 2001). There are no evidence indicating local mean sea level exceed the present level about 5600 cal yr BP (Martin et al., 2003). However, it is noteworthy that the level of studied sediments was more than 2 m above present sea level, and therefore, variations of less than that magnitude could not be recorded. Between 4560 and 4240 cal yr BP, an increase in sedimentation rate occurred at Juréia. However, by about 4320 cal yr BP, there is again a greater contribution of marine elements, a change in the behavior of isotopes and a decrease in content of some elements that indicate a lower terrestrial contribution and a sea level above the current one. The data corroborate the descriptions by Ybert et al. (2003); however, there is no indication of a sea level fall about 3470 cal yr BP at Juréia. This episode is possibly correlated with a 4.2 ka climate cycle identified in the North Atlantic (Bond et al., 1997, 2001).

Between 2120 and 1575 cal yr BP, there was a strong marine contribution in the paleolagoon reflected by an increase in the sedimentation rate, the enrichment of isotopes and a lower content of continental elements. These data agree with data suggesting a rising sea level about 2100 cal yr BP show by Martin et al. (2003). By 1575 cal yr BP, a greater continental contribution and a decrease in marine elements occurred in the paleolagoon, indicating a wetter period with a greater continental input. Mahiques et al. (2009) suggested that this contribution occurred about 1500 cal yr BP. The timing is broadly correlative with the last climate cycle identified in the North Atlantic by Bond et al. (1997, 2001), representing the Iron Age Cold Epoch. By 1300 cal yr BP, there is a great change in sediment source identified by lead isotopes; these data are consistent with a wetter period between 1300 and 675 cal yr BP (Ybert et al., 2003), an anomaly identified by Mahiques et al. (2009), and the beginning of the present conditions (Amaral et al., 2006). The establishment of

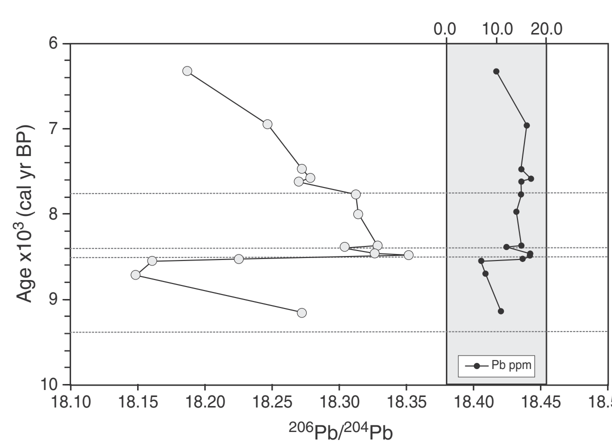

Figure 7. The pattern of the $^{206}Pb/^{204}Pb$ ratio of core S03 shows anomalies contemporaneous with records of the 8.2 ka event and abrupt changes between ~8500 and 7780 cal yr BP.

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the current conditions began about 300 cal yr BP and included the climatic deterioration associated with the Little Ice Age (Bond et al., 1997, 2001).

Evidence shows that environmental changes in the Northern Hemisphere commonly have been correlated with environmental in the Southern Hemisphere. The relationship is important for understanding global government debates and decisions regarding global climate changes and their local and regional environmental impacts. Many aspects of environmental responses to these global short-term events remain unclear as shown for the Juréia site in southeastern Brazil suggesting a need for new studies to improve the understanding of these complex environmental processes. Present results show that relatively small global climate changes can result in relatively large local and regional environmental responses.

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Appendix A. Supplementary data

Supplementary data to this article can be found online at doi:10.1016/j.yqres.2011.09.007.

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