The Holocene

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Abstract
We analyzed diatoms, lithology, and stable isotopes in a sediment core from the Ciénaga Grande de Santa Marta lagoon to reveal the history of late Holocene relative sea level rise and the ontogeny of the lagoon on the Caribbean coast of Colombia. At ~5300 cal. yr BP, the area was characterized by shallow, freshwater ponds that were prone to seasonal flooding. From ~4250 to ~2060 cal. yr BP, these ponds became brackish as relative sea level began to rise, and since ~2060 cal. yr BP, marine conditions have prevailed. In addition to tracking relative sea level rise, we also investigated periods of greater and lesser precipitation during times of brackish and marine conditions, respectively, as indicated by diatom-inferred changes in water salinity and shifts in the source of sediment organic matter. Comparisons with other regional paleoenvironmental records suggest that humid climate conditions prevailed in the Caribbean until about 4000 cal. yr BP. After that time, climate became more variable, with drier conditions registered from ~ 4000 to 2000 cal. yr BP. Wetter conditions returned after ~2000 cal. yr BP.

Introduction

Early Holocene global sea level rise has been studied intensively, and its causes and consequences are now fairly well understood (Cronin, 2012; Smith et al., 2011). There is consensus that the global effect of eustatic sea level rise during the last glacial–interglacial transition was felt until about 7000 cal. yr BP (Peltier, 2002; Smith et al., 2011), and that after that time, sea level stabilized until about 3000–2000 cal. yr BP, at which time it began to rise again (Cronin, 2012). This late Holocene sea level rise and its effects are difficult to identify, particularly in the Neotropics and in other areas far from the direct effects of melting ice caps (Martínez et al., 2010). Furthermore, different regions responded differently to sea level rise because of their particular physiographic settings and local tectonism (Smith, 1986), thereby confounding inferences about sea level rise and its causes.

Measurements of Holocene glacio-eustatic sea level rise in the Caribbean indicate a rapid rise of ~5 mm/yr between 8000 and 6000 cal. yr BP, and a lower rate of only ~0.25 mm/yr after 4000 cal. yr BP (Digerfeldt and Hendry, 1987; Feller et al., 1990; Rull et al., 1999). Late Holocene sea level rise and its effects on coastal environments has begun to receive considerable attention (Cronin, 2012; Woodroffe and Murray-Wallace, 2013), especially in light of anticipated rapid future sea level rise. In the Colombian Caribbean, several recent studies have evaluated the effects of sea level change on coastal erosion. In the western part of the

Colombian Caribbean, Correa et al. (2007) found coastal erosion rates of up to 7 mm/yr, with a consequent loss of pasture and agricultural lands that support dense human populations. On the Urabá Gulf, Bernal et al. (2005) found high sedimentation rates and decreasing water depths, which have negatively impacted navigation in the gulf. Paleontological and sedimentological studies of marine terraces near the coastal city of Cartagena suggest that the primary cause of the late Holocene sea level rise was local tectonics (Martínez et al., 2010). Van der Hammen and Noldus (1984) attribute the apparent sea level rise observed in the Ciénaga Grande de Santa Marta (CGSM) to local subsidence of the Magdalena River Delta. The origin of the CGSM lagoon is still

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under debate. According to Wiedemann (1973) and Cohen and Wiedemann (1973), the lagoon was formed by a relative sea level rise of about 2 m in the last 2300 years, inferred from radiocarbon dates and composition of peat samples from a north–south transect in the lagoon. The relative increase in sea level, however, could also have been a consequence of sediment compaction on the Magdalena River Delta (Van der Hammen and Noldus, 1984) or subsidence caused by local faulting, as suggested by Martínez et al. (2010). Jaramillo et al. (2012) argued that the lagoon formed by the interplay between mid-Holocene eustatic sea level rise and fluvial and sedimentological dynamics, generated by rivers draining the Sierra Nevada, east of the lagoon, and the Magdalena paleo-delta, west of the lagoon.

Most of what we know about late Holocene sea level changes in the Colombian Caribbean is based on pollen (Gonzalez et al., 2010; Palacios et al., 2012; Urrego et al., 2013; Van der Hammen and Noldus, 1984), petrographic analysis of organic matter (Cohen and Wiedemann, 1973), and sedimentological analyses (Jaramillo et al., 2012; Martínez et al., 2010). This study presents a record of sea level rise and environmental change for the Colombian Caribbean, inferred from analyses of diatoms, stable isotopes, lithology, and C/N ratios of organic matter, in a sediment core from the CGSM. Our main objective was to provide a comprehensive picture of late Holocene relative sea level change and environmental change in the southwestern Caribbean.

Study area

The CGSM is a large (450 km^2), brackish-water body in the northern lowlands of Colombia's Caribbean coast (Figure 1). The coastal lagoon is bordered to the west and south by lowland marshes that are part of the Magdalena River Delta, and on the east by the foothills of the Sierra Nevada de Santa Marta. To the north, it is partially isolated from the open ocean by a sand spit, the Isla de Salamanca, which possesses a 200- to 300-m-wide tidal channel at the northeast extreme of the lagoon (Figure 1). Mangrove forest grows in the surroundings, particularly in areas of high salinity (Rangel, 2012; Wiedemann, 1973). The lagoon is shallow, with water depths that range from 90 to 200 cm (Cohen and Wiedemann, 1973; Jaramillo et al., 2012). Water levels fluctuate by ~50 cm between the dry and rainy seasons. There is a salinity gradient across the water body, from marine (36‰) to brackish waters near the coastal areas in the north, to freshwater (0‰) in the south. The salinity gradient is maintained by dynamic interactions between marine water that enters through the tidal channel on the northeast end of the sand barrier, and freshwater inputs from the Magdalena River in the west, and the Frío, Sevilla, and Aracataca Rivers that originate in the Sierra Nevada de Santa Marta, to the east and southeast.

Salinity in the lagoon is higher during the dry season, when there is little input of freshwater, but lower during the rainy season, when there is abundant freshwater inflow (Van der Hammen and Noldus, 1984; Wiedemann, 1973). Salinity can fluctuate dramatically, as was demonstrated by the great freshwater flood event of 1969–1970, which could have been related to a very strong La Niña event (Gergis and Fowler, 2006). During those 2 years, the lagoon was dominated by freshwater, and as a consequence, oysters, which are adapted to brackish conditions or only short exposures to freshwater, died (Wiedemann, 1973). Cohen and Wiedemann (1973) and Wiedemann (1973) conducted petrographic and pollen analyses on ^{14}C-dated peat samples from 2 m below the sediment–water interface, at various sites in the lagoon (Figure 1). They concluded that the lagoon was an Everglades-type swamp that was inundated by a rapid marine transgression that flooded the marsh in <500 years, inferred from differences in composition and age of peats collected in the north and south parts of the lagoon, dated at ~2300 and ~1900 ^{14}C yr BP, respectively (Figure 1).
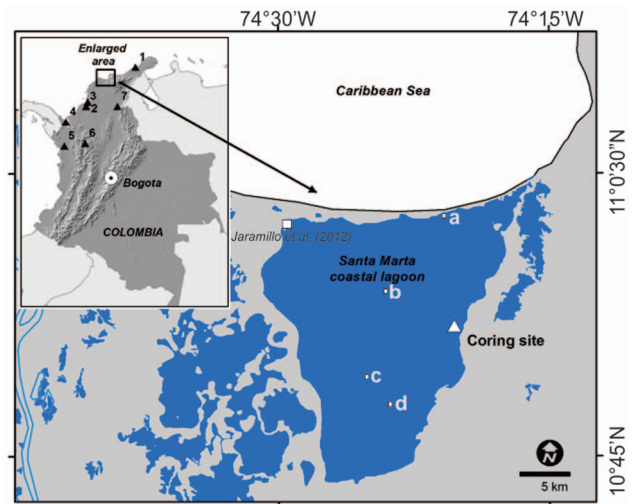

Figure 1. Map showing the location of the study site in the context of the Colombian Caribbean. The small map on the left shows Colombia, the study site and the location of other key sites mentioned in the text: (1) Guajira Peninsula, (2–3) marine terraces near Cartagena, (4) Cispatá Bay, (5) Atrato, (6) Cauca Valley, and (7) Lake Zapatosa. The enlarged map on the right shows the CGSM Lagoon, the marshes of the Magdalena River west of the lagoon, the site where the core was retrieved and the sections where previous ^{14}C dates have been obtained from the top of the peat: (a) 1930 ± 30 ^{14}C yr (Jaramillo et al., 2012), 2430 ± 85 ^{14}C yr; (b) 2300 ± 65 ^{14}C yr; (c) 1920 ± 65 ^{14}C yr; and (d) 1920 ± 65 ^{14}C yr (Wiedemann, 1973).
CGSM: Ciénaga Grande de Santa Marta.

We studied a sediment core collected in the east-central area of the CGSM, where the Frío River discharges into the coastal lagoon (Figure 1). We used diatoms to infer past salinity variations at the site, which reflect the dynamic balance between freshwater inputs from direct precipitation and river discharge, and marine input controlled by sea level change. We used the stable isotope signature of carbon to infer the source of sediment organic matter.

Methods

In October 2010, we collected a 6-m sediment core (Bocas de Lopez; 10°51′9.41″N, 74°19′56.18″W) from the CGSM using a Russian corer. Segments of 50 cm were sealed in PVC tubes and transported to the Paleoecology Laboratory (Instituto de Ciencias Naturales) of the Universidad Nacional de Colombia (Bogotá) for analyses. Working halves of each 50-cm core section were subsampled at 1-cm intervals for geochemical analysis and at 1-cm intervals every 5 cm for diatom analysis. Core chronology is based on six AMS dates, two on wood, one on plant material, and three on bulk organic sediment, determined at Beta Analytic, Inc. Radiocarbon ages were calibrated using IntCal09 (Reimer et al., 2011). This software was also used to calibrate radiocarbon dates from Wiedemann (1973). The age–depth model for our CGSM core was built by linear interpolation between dated depths (Figure 2).

Sediment samples were freeze-dried and crushed with a mortar and pestle prior to geochemical analysis. Total carbon and total nitrogen were measured using a Carlo Erba NA1500 CNS elemental analyzer. Carbonate carbon was determined by coulometric titration using an Automate acidification preparation device coupled with a UIC CO_2 coulometer. Percent organic carbon was calculated by subtraction of carbonate carbon from total carbon. Samples for stable isotope analysis of organic matter were treated with 2N HCl to remove carbonate

and then washed with distilled water to remove chloride. Approximately 50 mg of carbonate-free bulk sediment was loaded into tin sample capsules and placed in a 50-position automated carousel on the elemental analyzer. Combustion gases were carried in a helium stream through a Conflo II interface to a Finnigan-MAT 252 isotope ratio mass spectrometer. All carbon isotope results are expressed in standard delta notation relative to VPDB.

For diatom analysis, 0.1 g of dry sediment was placed in 30 mL of 30% H_2O_2 at room temperature for 48 h. Digested samples were brought to a volume of 30 mL with distilled water. Next, two 0.6–mL aliquots per sample were mounted on coverslips and dried at room temperature. Permanent slides were mounted in Zrax (RI ~1.7+). At least 400 diatom frustules were counted per slide. Other elements counted, but not included in the statistical analysis, were *Gamellodiscus* spores and silico-flagellates. Qualitative observations of charcoal size and abundance were also made. Diatom identification and ecology were based on Patrick and Reimer (1966), Krammer and Lange-Bertalot (1991, 1997), Torgan and Biancamano (1991), Moro and Fürstenberger (1997), Gaiser and Johansen (2000), Witkowski et al. (2000), Sanchez et al. (2003), and Torgan and Santos (2008). Diatom counts were transformed into percentages and plotted against depth using Tilia and Tiliagraph. Core zonation was done using CONISS (Grimm, 1987) within Tiliagraph.
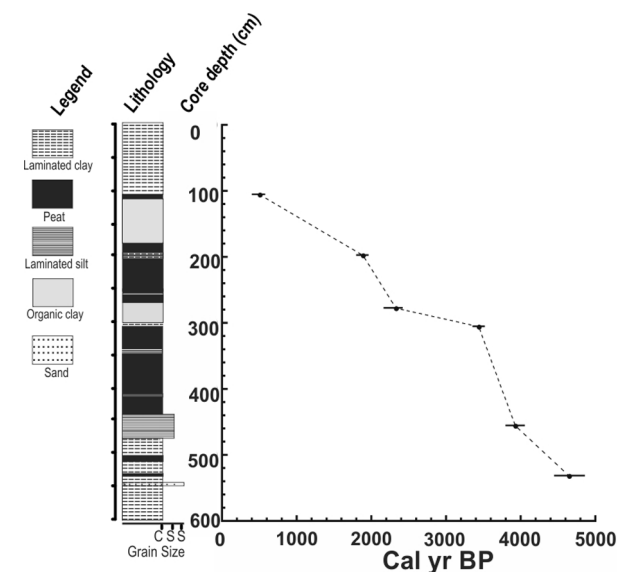

Figure 2. Stratigraphy and age model for the Bocas de Lopez core. Calibrated ages and error bars are presented.

Results

Core chronology and stratigraphy

Between 600 and 490 cm, the sediment is composed of laminated clay with peat (Figure 2), with an increase in sand content at 550–549 cm. From 490 to 413 cm, the sediment is composed of interbedded clay, silt, and peat. Fine laminations are present in the clay intervals. Between 486 and 458 cm, the grain size increases to predominantly silt, which is also laminated. From 413 to 300 cm, the sediment is mainly peat, locally laminated, sometimes with visible charcoal. From 300 to 248 cm, the sediment is composed of clay that varies from gray to black. A peat layer is present between 275 and 248 cm. Peat, with leaves, plant tissue, and calcareous shells is found from 246 to 184 cm. From 184 to 105 cm, the sediment is mainly organic-rich clay, and from 105 cm to the top, it is composed of laminated clay.

Eight radiocarbon dates from organic remains and organic-rich sediment were taken from different depths and lithologies in the core (Table 1). The basal date (600 cm) of 9190 ± 50 ^{14}C yr was not included in the age model because it was unusually old in light of the much younger ages up-core and there was no sedimentological, stratigraphical, or ichnological evidence of a hiatus. An additional argument for discarding this older date is that other water bodies in the Colombian Caribbean formed well after 9000 ^{14}C yr BP. For instance, such water bodies first appeared at 5482 ± 51 ^{14}C yr BP in Guajira (Urrego et al., 2013), northeast of the study site, and at 5905 ± 20 ^{14}C yr BP in the case of Lake El Totumo, west of the CGSM (Serna et al., personal communication). Dates immediately above, at 549 and 531 cm, are in reverse order. We accepted the date at 531 cm (4110 ± 50 ^{14}C yr BP), but rejected the date of 4030 ± 30 ^{14}C yr BP at 549 cm because the latter occurs in a lithologic interval with high sand content that could reflect a changing depositional environment. The date at 197 cm of 1930 ± 30 ^{14}C yr BP is potentially problematic because it came from a peat interval with carbonate shells and may have been subject to 'hard water error'. Nevertheless, this date is consistent with other dates for the CGSM reported by Wiedemann (1973) and Jaramillo et al. (2012). Wiedemann (1973) dated the top of the peat interval at ~2 m sediment depth in several cores from the lagoon. Two of these peat samples from cores taken in the south, not far from our study site (Figure 1), returned identical ages of 1920 ± 65 ^{14}C yr BP (1862 ± 70 cal. yr BP). Jaramillo et al. (2012) dated the top of the peat interval in a section collected from the north-west sector of the CGSM (Figure 1), which yielded the exact same age for the top of the peat in our core, that is, 1930 ± 30 ^{14}C yr BP. Ages for the top of the CGSM peat, obtained on previously studied cores from various areas of the lagoon, are all consistent and suggest that our date from 197 cm is reasonable.

Most of the lithologic units in the core are dated at their base and top (Figure 2), except for the bottom interval (600–550 cm) and topmost section of the core (upper 100 cm). Ages throughout the core were estimated using linear interpolation between dated

Table 1. Chronology of the Bocas de Lopez core.

Lab code	Name	Depth (cm)	Dated material	[14]C	Calibrated		σ	Median
					Min	Max		
320590	BOCALOPEZ105	105	Wood	430 ± 30	530	451	39	498
299970	BOCALOPEZ197	197	Sediment	1930 ± 30	1947	1820	36	1879
320591	BOCALOPEZ277	277	Wood	2290 ± 30	2353	2302	59	2322
320592	BOCALOPEZ305	305	Plant material	3220 ± 30	3485	3370	36	3431
299971	BOCALOPEZ455	455	Sediment	3610 ± 40	4000	3830	60	3920
299972	BOCALOPEZ531	531	Sediment	4110 ± 50	4824	4517	98	4641
320593[a]	BOCALOPEZ549	549	Wood	4030 ± 30	4572	4422	56	4485
296886[a]	BOCALOPEZ600	600	Sediment	9190 ± 50	10,445	10,241	72	10,352

[a]Samples not used in developing the core chronology (see text for explanation).

depths. The age at the bottom of the section was estimated by extrapolation of the linear sedimentation rate between 455 and 531 cm to the bottom of the core, which yielded a basal date for the CGSM of ~5300 cal. yr BP. Although this date is tentative, it is consistent with the timing of formation of other water bodies in the Colombian Caribbean, as mentioned above.

Diatom analysis

Three main diatom zones were identified using cluster analysis and visual inspection of the stratigraphic distribution of diatoms (Figure 3a and b). All species with <2% representation were deleted from analyses. The average percentage for each taxon is presented in parentheses.

Zone 1 (600–490 cm). This zone is largely dominated by benthic and freshwater species *Diadesmis confervacea* (mean = 17%, with a peak at 550 cm of 53%) and *Eunotia monodon* (mean = 11%, with a peak at 515 cm of 17%), but also has substantial numbers of planktonic *Aulacoseira italica* (mean = 11%, with peaks of 40% at 560 and 540 cm). Planktonic species *Aulacoseira granulata* (mean = 4% with a peak of 20% at 595 cm) and *Aulacoseira* cf. *distans* (one peak of 18% at 585 cm) are more abundant at the bottom of the zone, whereas *Cyclotella meneghiniana* is more abundant (mean = 9%) in the middle, and *Aulacoseira herzogii* at the top (mean = 7%). Species of freshwater genera *Eunotia* and *Pinnularia* have their highest abundance in this zone. Epiphytic species *Cocconeis placentula* and *Cymbella silesiaca* are also present throughout the zone. Saline-tolerant species *C. meneghiniana* and *Pseudostaurosira subsalina* are present, with an increase in the middle of the zone.

Zone 2 (490–230 cm). This interval is marked by variable preservation of valves. At 440 cm, 400–355 cm, 340 cm, and 325–300 cm, diatom numbers were insufficient to reach the minimum count. Some samples show signs of valve dissolution, whereas other samples possess fractured valves (e.g. at 405 cm). In samples from 440 cm and 400–390 cm, other siliceous microfossils such as phytoliths are present, as are pieces of plant tissue and charcoal fragments of considerable size (100 μ). At depths where diatoms are present, the assemblage is dominated by saline-tolerant species *P. subsalina* (40%) and *C. meneghiniana* (9%). Freshwater benthic species are also present: *D. confervacea* (mean = 7%, with a peak of 20% at the bottom of the zone), *Nitzschia amphibia* (mean = 9%, with peaks of 14% at 485, 350, 345, and 295 cm), *Cyclotella stelligera* (mean = 3%) and *C. placentula* (mean = 2.5%). *Pinnularia maior* displays a peak (18%) at the bottom of the zone (430 cm). Brackish-water and marine species that represent minor proportions include *Achnanthes submarina, Campylodiscus* sp., *Chaetoceros* aff. *Ceratosporus*, and *Nitzschia scalaris*. Marine diatom *Amphora marina* and brackish-marine *Opephora guenter-grassii* peak at the top of the zone, with 33% and 22%, respectively. *A. marina* remains are represented mainly by the central areas of the valves. At depths from 270 to 265 cm, marine, saline-tolerant and freshwater species co-dominate: *Actinocyclus curvatulus* (10%), *Actinocyclus normanii* (4%), *C. meneghiniana* (13%), *C. placentula* (7%), and *D. confervacea* (6%). *A. italica*, although present in smaller percentages compared with the previous zone, is still present (5%). *Gamellodiscus* spp. spores and silico-flagellates peak in this zone (Figure 3b).

Zone 3 (230–0 cm). This zone is dominated by the marine and brackish planktonic species *A. curvatulus* (42%) and *A. normanii* (12%), respectively. *A. normanii* tolerates a wide range of salt concentrations and is reported as both a freshwater (Juggins, 2013) and brackish-water diatom (Witkowski et al., 2000).

This species inhabits the CGSM today and is considered a brackish-water species that prefers waters with relatively low salinity and high silica content (Sanchez et al., 2003). We therefore consider it a brackish-water diatom. Other saline diatoms present are *Nitzschia compressa* var. *compressa* (5%) and *Caloneis amphisbaena* (4%). Brackish-water *O. guenter-grassii* peaks at the top of the zone, with 55%. Freshwater benthic and planktonic species *A. herzogii, A. italica, E. monodon*, and *D. confervacea* increase up to ~20% at 145, 105–90, 55, and 15 cm. The samples from 225 to 205 cm possessed plant tissue, but did not contain enough diatoms to reach the minimum count.

Geochemistry

Minimum C/N values ~6 are registered between about 600 and 550 cm and values >20 are recorded between about 550 and 525 cm. Values of C/N ~12 are recorded between about 525 and 490 cm (Figure 4). The largest fluctuations in C/N are registered between 490 and 195 cm. Within this interval, all C/N values are >12, with values > 20 between about 400 and 350 cm, 330 and 300 cm, and maxima > 40 between about 250 and 200 cm. Low values of C/N (~10) are again recorded between about 200 and 100 cm, with a slight increase in C/N values to 15–20 between 100 and 0 cm.

Relatively high δ^{13}C values, between −26‰ and −24‰, coincide with the lowest C/N values (~6) between about 600 and 550 cm. A decrease in δ^{13}C is recorded between about 550 and 490 cm, and δ^{13}C values of about −28‰ are registered between 490 and 300 cm. A rapid decline in δ^{13}C is recorded around 300 cm, with values around −29‰ occurring between 360 and 200 cm. Two periods of high δ^{13}C, with a peak of −20.4‰ at about 150 cm and a low of −28.4‰ at 140 cm, are recorded between 190 and 150 cm and between 125 and 105 cm. The δ^{13}C values generally increase, but display high variability from 100 cm depth (−29‰) to the core top (−26.1‰).

Interpretation of results

Diatoms were grouped into four categories according to salinity preference: freshwater, freshwater but saline-tolerant, brackish-water, and marine (Figure 3b). A 20% increase in freshwater species in a marine-dominated assemblage was interpreted as indicative of a salinity decrease. We interpret C/N values between 4 and 10 as indicative of organic matter coming primarily from algae (Kendall et al., 2001; Lamb et al., 2006; Meyers, 1994). C/N values > 12 are interpreted as indicating organic matter contribution from vascular plants, given their high lignin and cellulose content, and their relatively low amount of nitrogen (Lamb et al., 2006; Prahl et al., 1980). Values of δ^{13}C in our core range mainly between −23‰ and −30‰, indicative of being derived from C3 terrestrial plants (range −23‰ to −34‰; Lamb et al., 2006; Smith and Epstein, 1971). Only samples from 193, 153, and 149 cm have δ^{13}C greater than −23‰, which indicates a phytoplankton origin (range −20‰ to 23‰; Smith and Epstein, 1971). Higher C/N values correspond to peat intervals in the sediment, and C/N ratios > 20 reflect periods of greater terrestrial vegetation input. Such 'terrestrialization' occurred at times when water levels decreased and terrestrial vegetation expanded. Diatom analysis enabled the identification of three main environmental periods as follows.

Period 1 (~5300–4250 cal. yr BP; diatom zone 1)

Diatoms in this zone prefer shallow, freshwater, lentic environments of low pH and low conductivity. Thus, we infer that during this period, shallow, freshwater ponds predominated on the landscape. Periods of higher water levels and increased turbulence also occurred during this time, as indicated by the increasing
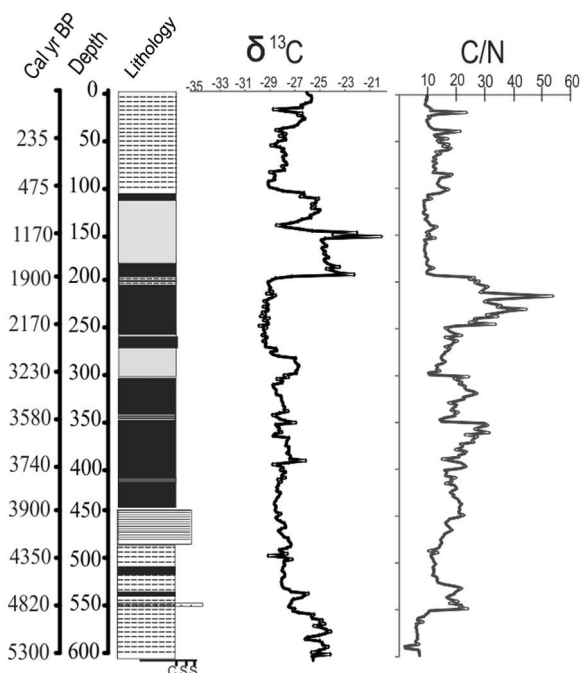
Figure 3. Diatom composition of the Bocas de Lopez core. Core intervals where diatoms are absent are indicated. (a) Diatom zones and ecological groups of the diatoms, according to salinity preferences. (b) Lithology of the core (modified from Jaramillo et al., 2012), and main groups of diatoms, with indications of the number of cells of other fossils such as marine diatom spores, silico-flagellates, and presence of bivalves and wood fragments.

Figure 4. Lithology and stable isotope composition (δ^{13}C) and C/N of organic matter in the Bocas de Lopez core.

proportions of planktonic *Aulacoseira* species. At the beginning of this period, macrophytes were abundant, as indicated by the large numbers of epiphytic diatoms *C. placentula* and *C. silesiaca*.

Lithology indicates sediment accumulation under quiescent water conditions, but subject to seasonal flooding, as indicated by laminations and occasional presence of sand. At 550 cm (~4820 cal. yr BP), total C increases and C/N changes from <10 to >20, suggesting an increase in organic matter accumulation and possibly a shift in the relative proportions of organic matter sources, from predominantly algae to predominantly terrestrial plants. The sediment becomes increasingly rich in sand and charcoal and the soil diatom *D. confervacea* peaks. This is interpreted as a change from largely aquatic to mainly terrestrial conditions. The slight increase in saline-tolerant diatoms *P. subsalina* and *C. meneghiniana* suggests that during this low water period, salinity increased, indicating a decrease in freshwater input. Nevertheless, an overall freshwater setting is supported by the pollen record (Van der Hammen and Noldus, 1984; Figure 5), which shows abundant marsh elements such as Cyperaceae, Poaceae, *Typha*, Chenopodiaceae, and aquatic *Monolete psilate* and *Trilete psilate* plants and a surrounding flooded forest with *Alchornea*, *Acalypha*, Bombacaceae, Guttiferae, Lecythidaceae, and minor mangrove elements. According to these authors, the area at that time was still part of the Magdalena River Delta.

Period 2 (~4250–2060 cal. yr BP; diatom zone 2)

Variations in the preservation and composition of diatoms in this period suggest frequent fluctuations in water level and physicochemical characteristics of the water. Compared with the previous zone, water became brackish, as indicated by the increase in saline-tolerant and brackish species. During this period, marine incursions occurred *c.* 2255 cal. yr BP (265 cm) and from ~2145 to 2120 cal. yr BP (245–240 cm), as indicated by high numbers of spores of marine *Gamellodiscus* spp. and an increase in silicoflagellates and marine diatoms *A. marina* and *A. curvatulus*. Despite the marine incursions, periods of fresher water, inferred from increases in freshwater, soil, and epiphytic diatoms, occurred from ~3870 to 3840 (440–430 cm) and at ~3580 (350 cm) cal. yr

BP. Periods of terrestrialization and the occurrence of nearby fires are indicated by the poor preservation of diatoms, the presence of phytoliths and charcoal, and a positive correlation between peaks in C/N and peat layers. These are dated at 3740–3595 (400–355 cm) and 3495–3230 (325–300 cm) cal. yr BP. The lowest sedimentation rate in the core, ~0.025 cm/yr, is recorded between 305 and 277 cm. Oscillations in total C, along with C/N values > 10 and lithologic changes from silt to peat to clay, indicate rapid shifts in the relative sources of organic matter. These, in turn, are interpreted as reflecting rapid oscillations in water level, from very low stage, accompanied by vegetation expansion, to periods of higher water level associated with substantial hydrologic energy. According to Van der Hammen and Noldus (1984), freshwater herbs expanded at the expense of forest, and so did the red mangrove (*Rhizophora*) and Chenopodiaceae (Figure 5). The pollen record also shows that there were periods when mangrove vegetation declined, most likely because of a reduction in water salinity, and when freshwater aquatic vegetation like *Sagittaria* and *Typha* increased, indicative of expansion of the lagoon. Pollen, diatoms, lithology, and isotopes all point to a net increase in marine influence, which created a brackish environment. Within this period, times of increased precipitation alternated with times of reduced precipitation, that is, dry conditions.

Period 3 (~2060 cal. yr BP to present; diatom zone 3)

Marine influence is indicated by the diatom composition of zone 3, dominated by marine *A. curvatulus* and brackish *A. normanii*. The onset of high-salinity conditions and thus marine influence is also indicated by the change in the sediment from peat to detrital-rich clay and laminated clay. This lithologic change is dated at 1930 ± 30 ^{14}C yr (1990 cal. yr BP, 1879 cal. yr BP) in the northwest part of the CGSM (Jaramillo et al., 2012; Figure 1) and at 1920 ± 65 ^{14}C yr (1862 ± 70 cal. yr BP) in the southern parts of the lagoon by Wiedemann (1973) and in this study (Figure 1). Slightly older dates on the lithological shift, measured in cores from the northeastern and central parts of the lagoon (2430 ± 85 ^{14}C yr (2484 ± 148 cal. yr BP) and 2300 ± 65 ^{14}C yr (2318 ± 69 cal. yr BP; Figure 1) suggest that the marine transgression started in the north, but reached the south and north-west in a matter of several centuries, forming the lagoon environment we know today. The high-salinity conditions could have been caused by isolation of the lagoon from the ocean when the sand spit formed. Similar to the previous period, there were frequent environmental changes. Freshwater pulses, resulting in reduced salinity, are indicated by increases in brackish-water diatoms at ~1025 cal. yr BP (140 cm) and in freshwater species at 450 (95 cm), 260 (55 cm), and 70 (15 cm) cal. yr BP. A period of terrestrialization is indicated by the poor preservation of diatoms, peaks in C/N, minima in carbon isotopes, and the formation of peat around 2035–1895 cal. yr BP (225–200 cm).

Marine influence is also inferred from the carbon isotopes. There is a sharp increase in δ^{13}C values at ~1820 cal. yr BP (193 cm) that coincides with a sharp increase in the relative percentage of marine diatoms. One control on phytoplankton δ^{13}C is carbon source. Dissolved CO_2 in equilibrium with the atmosphere has a δ^{13}C value of ~ −8‰, whereas HCO_3^- has a δ^{13}C value near 0‰ (Keeley and Sandquist, 1992). There is relatively less CO_2 in the marine environment than in most fresh waters because the ratio of CO_2 to HCO_3^- is a function of water pH (the higher the pH, the lower the ratio CO_2/HCO_3^-). Thus, marine algae should have higher δ^{13}C values than algae from most freshwater systems.

The pollen data of Van der Hammen and Noldus (1984) revealed that during Period 3, the mangrove forest was composed mainly of red (*Rhizophora mangle*), with minor presence of black mangrove (*Avicennia germinans*; Figure 5). Petrographic study of peat from the south-central area of the lagoon and dated to 1920 ±
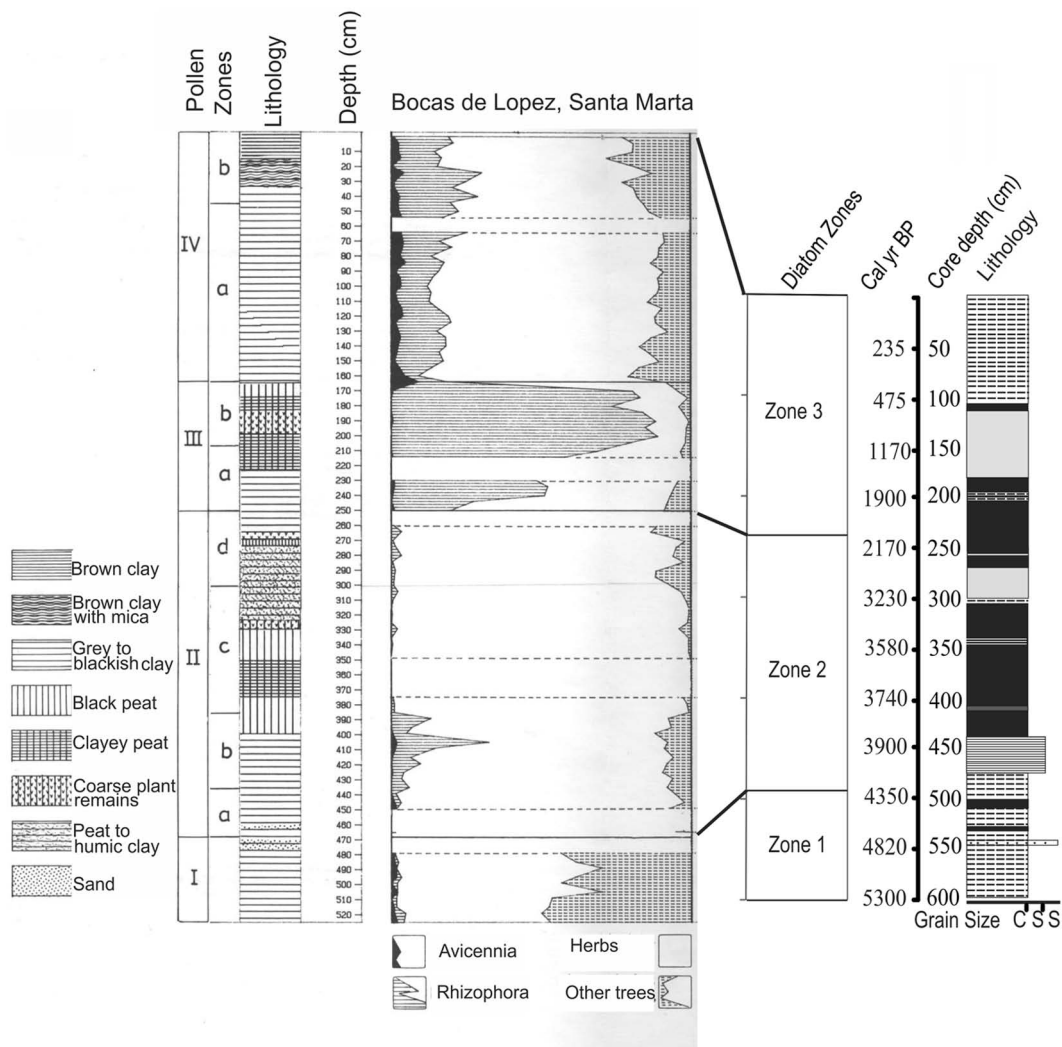
Figure 5. Summary of the pollen diagram presented in Van der Hammen and Noldus (1984; left) and correlation with results from this study (right). Figure on the left was reproduced with permission from J Cramer (www.schweizerbart.de).

65 ¹⁴C (1862 ± 70 cal. yr BP; Wiedemann, 1973), indicates it was mangrove-derived, which was interpreted by the authors as evidence for expansion of mangrove forest. This also indicates stronger marine influence. Surrounding the lagoon, a forest and herb vegetation grew at times, as did palm trees, the latter interpreted by Van der Hammen and Noldus (1984) as indicative of local human activity.

Regional environmental and climatic reconstruction

Between ~5300 and 4250 cal. yr BP, the CGSM coastal lagoon did not exist. Instead, the area was covered by freshwater ponds that flooded seasonally, which we interpret as indicating prevailing humid conditions. A brief period of drier climate occurred ~4820 cal. yr BP, during which the freshwater supply decreased and terrestrial vegetation expanded as water levels fell. Humid conditions are also inferred for northeastern Colombia for the period ~4700 and 4000 cal. yr BP. Lake Zapatosa, south of the CGSM lagoon (Figures 1 and 6), had high water levels (Romero and Rangel, 2013) as did the high-altitude lakes of the Sierra Nevada de Santa Marta, east of the study area (Van der Hammen, 1984). Humid conditions are also inferred for the Guajira Peninsula, northeast of the CGSM (Figures 1 and 6), from development of a mangrove forest that expanded under a delicate balance between increased tidal influence, a consequence of sea level rise, and prevailing humid conditions (Urrego et al., 2013). Elsewhere

around the Caribbean, in such places as the Cariaco Basin, north of Venezuela (Haug et al., 2001), and Lake Miragoane, Haiti (Hodell et al., 1991), humid conditions are also inferred for this time. Mayewski et al. (2004) proposed that generally humid climates characterized the entire Caribbean during that period.

By ~4250 cal. yr BP, the CGSM coastal lagoon began to form as a consequence of relative sea level rise. Within this period of brackish conditions, variable climate conditions prevailed. There were several short episodes of reduced salinity, interpreted as having been caused by increased precipitation, which alternated with times of terrestrialization, reflecting reduced precipitation. Other records from nearby areas also show variable conditions. Between ~4000 and ~3000 cal. yr BP, *Rhizophora* mangrove forest was well established in the Guajira region, as sea level stabilized and humid conditions persisted (Urrego et al., 2013). This, however, changed, and from ~3000 until 2600 cal. yr BP, the more saline-tolerant *Avicennia* took over. This shift is interpreted by the authors as reflecting an increase in salinity because of reduced rainfall. High water levels, and thus humid conditions, prevailed in Lake Zapatosa, south of CGSM (Figures 1 and 6), from ~5500 to 4100 cal. yr BP and from ~3500 until ~2000 cal. yr BP, whereas low water levels, and thus dry climates, reigned from 4100 to ~3500 cal. yr BP (Romero and Rangel, 2013). To the west, near Cartagena (Figure 1), large amounts of sediment discharge into the Caribbean occurred between ~3500 and 3000 cal. yr BP, and a pronounced reduction in sediment load was
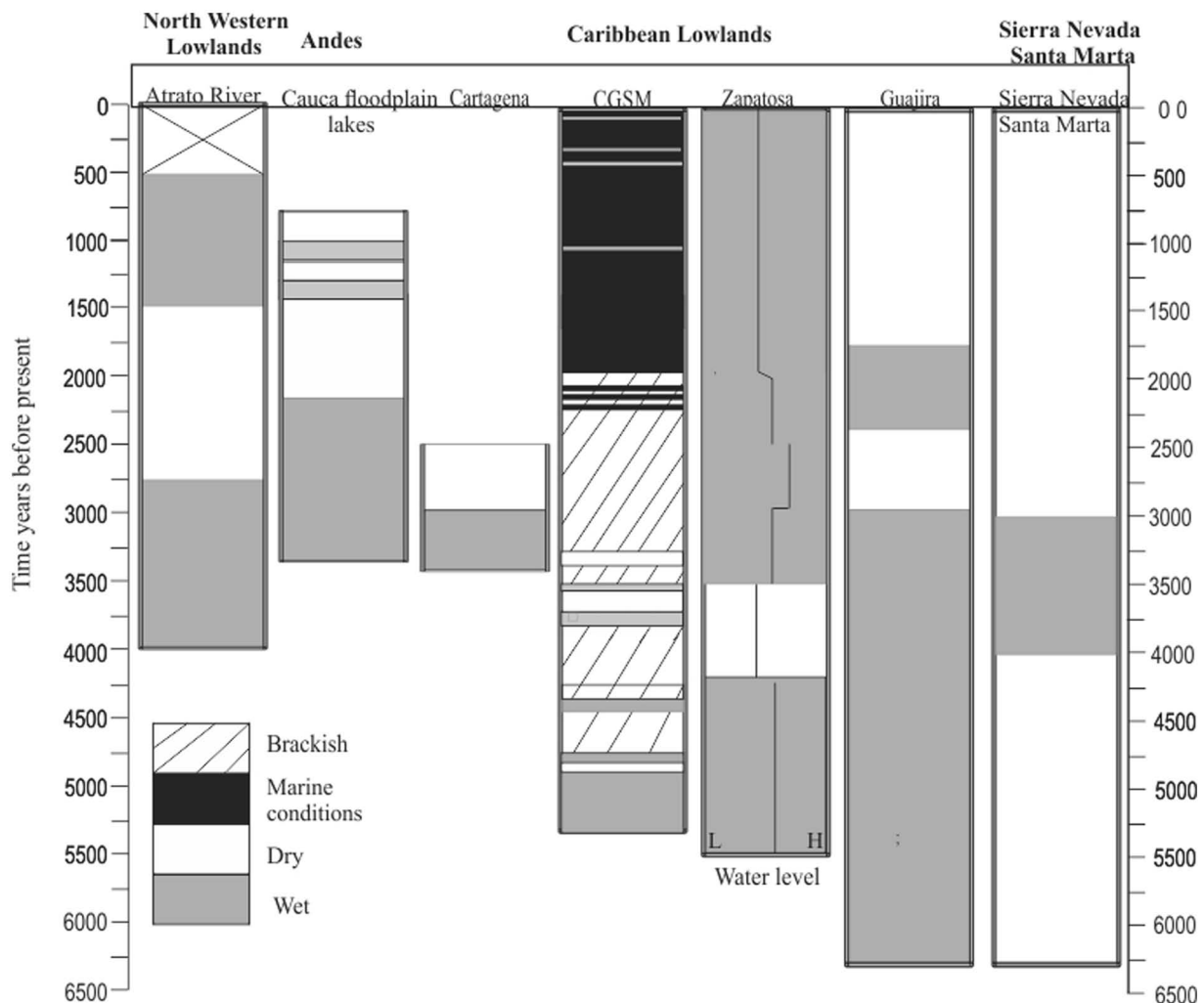
Figure 6. Summary of other paleoenvironmental and paleoclimate records from the Colombian Caribbean mentioned in the text.

identified between ~3000 and 2500 cal. yr BP. This reduction was interpreted as indicative of a switch from humid to dry conditions (Martínez et al., 2010). Middle Holocene drying is also reported farther north, in the lowlands of Petén, Guatemala, where the reduction in rainfall began around 4500 cal. yr BP (Mueller et al., 2009). By ~3000 cal. yr BP, the drying trend had reversed somewhat. Slightly farther north, in the central Yucatán Peninsula of Mexico, drier times were registered between about 3000 and 1000 cal. yr BP (Hodell et al., 1991, 1995), with episodic strong droughts occurring about every ~210 years (Hodell et al., 2001).

Southwest of the CGSM, in the Chocó region of Colombia, humid conditions, as indicated by an increase in flood vegetation in the Atrato floodplain (Figures 1 and 6), prevailed from ~4000 to ~2800 cal. yr BP. After that time and until ~1500 cal. yr BP, drier climates were established, as indicated by the reduction in flood vegetation. Thereafter, humid conditions returned (Urrego et al., 2006).

The change to drier climate in northern Colombia between ~3000 and ~2500 cal. yr BP was most likely linked to the more southerly migration of the ITCZ (Haug et al., 2001), which reduced rainfall in the northern regions of Colombia, but increased it in the southern regions and in Ecuador (Gonzalez-Carranza et al., 2012; Mollier-Vogel et al., 2013; Velez et al., 2003). In the Cauca Valley of Colombia, humid conditions prevailed from ~3500 to ~2400 cal. yr BP, inferred from increased effect of flood pulses on the Cauca River floodplain lakes (Velez et al., 2013). This increased flooding, that is, hydrologic shift, was also recorded by the shift from *igapó-* to *várcea-*type lakes in the

Cauca region (Martínez et al., 2010). From ~2400 to ~800 cal. yr BP, fluctuations in the frequency of river pulses and periods of pedogenesis indicate variable conditions for the northern part of the Cauca Valley.

Using mainly pollen records, Marchant and Hooghiemstra (2004) identified a major shift to wetter conditions in northern South America centered at 4000 cal. yr BP. The authors proposed that the development of the El Niño Southern Oscillation (ENSO), changes in solar variation and volcanic activity, oceanic circulation, and local conditions were among possible causes for this change. High variability between wet and dry climate after ~4100 cal. yr BP was most likely related to the increase in ENSO activity that occurred between ~5600 and 3300 cal. yr BP, with longer La Niña events between 4400 and 3600 cal. yr BP (Wanner et al., 2008). Mayewski et al. (2004) found that rainfall over the Caribbean became more erratic from ~3500 to ~2500 cal. yr BP.

Since 2060 cal. yr BP, the CGSM lagoon has been under permanent marine influence, most likely a consequence of relative sea level rise. Seasonality continued to influence hydrologic conditions, as in the previous period; however, an overall increase in humidity is indicated by the return of forest, frequent freshwater influx and less terrestrialization. Marked increases in seasonality are evident from the frequent vegetation changes in the Guajira region and the frequent hydrological changes recorded by the Cauca floodplain lakes. The humidity increase recorded ~2000 yr BP, inferred from our CGSM record and in records from Venezuela (Polissar et al., 2013), could be related to the increase in La Niña activity reported by Wanner et al. (2008) and Gergis and Fowler (2006).

Sea level rise is first identified at ~6200 cal. yr BP in the Guajira region, northeast Colombia, then at ~4100 cal. yr BP in the CGSM, and finally in Cispatá Bay at ~1075 cal. yr BP (Palacíos et al., 2012) to 900 cal. yr BP (Castaño et al., 2009), southwest of the CGSM lagoon (Figure 1). The causes of sea level change are beyond the scope of this study, but our record suggests that sea level rise in the CGSM reflects dynamics between fluvial input from inland rivers and increasing eustatic sea level in the late Holocene (Cronin, 2012).

Conclusion

Increasing relative sea levels are first recorded in Colombia's CGSM coastal lagoon after ~4250 cal. yr BP, when freshwater ponds became brackish. Since ~2060 cal. yr BP, decidedly marine conditions have prevailed. Our date for the establishment of marine conditions (~2060 cal. yr BP) agrees with the date reported by Wiedemann (1973) for the central-south and southern part of the lagoon (~1862 ± 70 cal. yr BP) that indicates that the transgression was rapid.

Around 5000–4000 yr BP, several water bodies, including Arcial and Atrato, as well as CGSM and El Totumo, formed in northern Colombia. During that time, stable humid conditions prevailed. After ~ 4000 cal. yr BP, climate became more variable. Variable, but overall drier conditions from ~4000 to 2000 cal. yr BP reflect lower precipitation, as the ITCZ migrated farther southward. Relatively wetter conditions after ~2000 cal. yr BP were likely related to intensification of the La Niña phase of the ENSO phenomenon. Future work in the northern and western parts of the lagoon is needed to clarify the pace of transgression, the age of sandbar formation, and its effect on the pre-lagoonal environment.

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