A molecular evaluation of bulk organic carbon-isotope chemostratigraphy for terrestrial correlations: An example from two Paleocene–Eocene tropical sequences

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Abstract

The dynamics associated with the carbon cycle and the linkage between the oceans, the atmosphere, and land plants provide an opportunity to correlate marine and terrestrial sedimentary sequences using stable isotopes of carbon (δ13C), but few studies have tested this approach. For instance, it has been proposed that changes in plant community (e.g., gymnosperm-dominated vs. angiosperm-dominated) could have significantly altered/amplified the carbon-isotope ratios of bulk sedimentary organic matter derived from land plants (δ13CTOM), compared to that of the marine carbonates (δ13Ccarbonate). Here, δ13CTOM values in a terrestrial sequence of the Colombian tropics are compared to the composite Paleocene–Eocene δ13Ccarbonate curve from Zachos et al. (2001) to evaluate the use of δ13CTOM values as a reliable chronostratigraphic tool. Sediments of the studied terrestrial sequences were deposited in a transitional setting dominated by mudstones, coals, and small lenses of sandstones (Late Cretaceous–Middle Paleocene sediments) and in a mixture of deltaic and fluvial conditions (Late Paleocene–Early Eocene sediments). The biostratigraphic control was based on palynological zones for the region. The δ13CTOM values for the studied terrestrial sequence show three carbon-isotope excursions, which correlate closely with those present in the marine record. The δ13CTOM values decreased from −24.2‰ to −26.5‰ in sediments accumulated during Early to Middle Paleocene, increased from −26.5‰ to −23.8‰ during the Late Paleocene, and decreased from −23.8‰ to −26.5‰ near the Paleocene–Eocene boundary (52–50 Ma). Selected biomarkers indicate that most of the organic matter derived from both gymnosperms and angiosperms. Moreover, the analyses of selected biomarker ratios (CPI, Pr/Ph, Tm, and δ[4]/[1] hopanes) show some diagenetic transformation in the preserved organic matter. However, no correlation between diagenesis and δ13CTOM Values was detected, thus suggesting that δ13CTOM could be correlated with δ13Ccarbonate values. The close correspondence that was found between δ13CTOM and δ13Ccarbonate values (Δδ13CTOM–carbonate ~ −27‰) provides support to the hypothesis that a tight land-plant–oceans linkage exists through geologic timescales via atmospheric carbon dioxide.

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1. Introduction

Significant chemical, physical, and biological changes in both marine and terrestrial environments have affected the carbon cycle during the Phanerozoic (e.g., Arens and Jahren, 2000; Beerling et al., 2001; Freeman and Colarussos, 2001; Hesselbo et al., 2002; Thomas et al., 2002; Hesselbo et al., 2003; Jahren et al., 2005; Jaramillo et al., 2006; Kaiser et al., 2006; Smith et al., 2007) by affecting the magnitude of carbon fluxes between different carbon reservoirs over geologic time scales and by altering the partitioning of carbon isotopes. If these perturbations of the carbon cycle are global in scale, they could offer an opportunity to correlate marine and terrestrial sedimentary sequences, which at the moment is limited to a few proxies. Some studies have in fact proposed the use of stable isotopes of carbon as a chronostratigraphic tool for marine and terrestrial sequences (e.g., Scholle and Arthur, 1980; Hasegawa, 1997; Arens and Jahren, 2000; Strauss and Peters-Kottig, 2003; Hesselbo et al., 2007). These studies rely on two key assumptions: (1) both δ13Ccarbonate values and δ13C values of organic matter derived from higher plants reflect parallel changes in the carbon isotopic composition of the marine and terrestrial realms, and (2) measured δ13C values are identical to those of the pristine materials.

The first assumption is based on the notion that isotopic equilibrium is reached within the different reservoirs over geologic timescales (i.e., differences in isotopic values remain constant after equilibration) (e.g., Gröcke, 2002; Strauss and Peters-Kottig, 2003). Despite the importance of the secular changes over long-term time scales (>10⁶ yrs), significant partitioning of carbon isotopes occurs over short-term timescales, which ultimately drives the isotopic...
composition of both marine carbonates ($\delta^{13}C_{\text{carbonate}}$) and terrestrial plants ($\delta^{13}C_{\text{plant}}$).

The second assumption has been challenged by several studies showing that carbon isotopic composition of bulk terrestrial organic matter ($\delta^{13}C_{\text{TOM}}$) may not faithfully reflect pristine $\delta^{13}C_{\text{plant}}$ values (e.g., Gröcke, 1998, 2002; Bergen and Poole, 2002) due to the preferential preservation of different plant components during diagenesis (e.g., Beerling and Royer, 2002). These plant components have a different isotopic composition relative to that of whole plant tissues (Bergen and Poole, 2002), and their carbon isotopic composition varies as a result of the different isotope fractionation effects occurring during biosynthesis (Galimov, 1985; Farquhar et al., 1989; Hayes, 2001; Gröcke, 2002). In addition, aerobic bacteria preferentially metabolize $^{12}C$-enriched organic matter during diagenesis, potentially enriching the remaining organic compounds in $^{13}C$ and possibly making $\delta^{13}C_{\text{TOM}}$ values more positive relative to $\delta^{13}C_{\text{plant}}$ values (Hartgers et al., 1994; Bergen and Poole, 2002). To date, only a few studies (e.g., Bergen and Poole, 2002; Poole et al., 2004) have tested the effects of diagenesis on $\delta^{13}C$ values from fossilized plant parts (e.g., wood) using reliable tools (e.g., molecular biomarkers). For that reason, the goal of this study is to test whether $\delta^{13}C_{\text{TOM}}$ values mirror the isotopic shifts that are recorded in marine carbonates during coeval time intervals. Moreover, the effect of diagenetic processes and differences in the source of organic matter are evaluated to infer their effect on $\delta^{13}C_{\text{TOM}}$ values. To achieve these goals, the Lower Paleogene marine carbon-isotope record (Zachos et al., 2001) is compared to that of a terrestrial sequence from the South American tropics. This chronostratigraphic interval was selected for the study because several isotopic anomalies have been reported in several marine sequences accumulated during the Early Paleogene (e.g., Koch et al., 1992; Veizer et al., 1999; Zachos et al., 2001; Hollis et al., 2005), thus providing an excellent reference for comparing $\delta^{13}C_{\text{carbonate}}$ and $\delta^{13}C_{\text{TOM}}$ values. The effect of diagenesis on $\delta^{13}C_{\text{TOM}}$ values is evaluated through the use of geochemical biomarkers. Biomarker parameters are also used qualitatively to evaluate microbial- vs. plant-derived contributions to organic matter, thereby allowing an evaluation of the potential effect of microbial degradation on $\delta^{13}C_{\text{TOM}}$ values. These results will then permit us to determine whether $\delta^{13}C_{\text{TOM}}$ values are a reliable proxy for chronostratigraphic and paleoclimatic studies.

2. Geological setting and sampling site

The samples for this study were collected from well sites located in two contiguous basins in northern South America. Site-A (8°12′ N, 72°1′ W) is located within the Catatumbo basin, whereas Core-B (9°34′16″ N, 73°16′45″ W) is located within the Cesar–Rancheria basin (Fig. 1). Sampling was performed at about 10 m intervals on each site to cover the desired time span (Latest Cretaceous–Early Eocene), yielding a set of 134 samples (69 from the Site-A and 65 from the Core-B wells). The time span encompassed by the samples was chronostratigraphically constrained with the aid of pollen biozones, previously calibrated with the marine
chronostratigraphic record (Jaramillo and Rueda, 2004; Jaramillo et al., 2005) (Fig. 2). The average sedimentation rates were estimated on the basis of the ages provided by the pollen biozones for each sedimentary basin and the thickness of each biozone in the wells. For the sampled interval, the average sedimentation rates were similar for both sequences (41.36 m/M.y. for Site-A and 41.44 m/M.y. for Core-B). These sampled sediments consisted of mudstones, coals, and small lenses of sandstones probably accumulated in a transitional setting from the Late Cretaceous to the Middle Paleocene (Catatumbo and Barco Formations) and in a mixture of deltaic and fluvial settings, possibly including oxbow lake environments, that existed from Late Paleocene to Early Eocene (Cuervos Formation).

3. Analytical methods

3.1. Bulk sediment isotopic analysis

Stable carbon-isotope values of bulk sediment (δ13CTOM) were measured via flash-pyrolysis at 1100 °C in a Costech elemental analyzer fitted to a Thermo Finnigan Delta plusXL isotope ratio mass spectrometer (Department of Geological and Atmospheric Sciences at Iowa State University). Carbonate was removed from the samples by HCl digestion. Analytical precision and accuracy were determined on the basis of repeated analysis of two internal lab standards calibrated against the internationally accepted V-PDB standard. Overall uncertainty was better than 0.08‰. Organic carbon content was determined on the basis of the liberated CO2 in the elemental analyzer. To minimize noise in the obtained data, a five-point moving average was applied to the δ13CTOM values for a better comparison between δ13Ccarbonate and δ13CTOM. Comparisons between δ13CTOM and other geochemical parameters were performed using original, non-smoothed isotopic values.

3.2. Biomarker analysis

A set of 27 samples from the initial 134 were selected for biomarker analysis on the basis of their organic carbon content and on their relative stratigraphic location to represent the entire sequence. Soluble organic matter (SOM) extractions followed the methodology proposed by Otto et al. (2005). The saturated and aromatic fractions were separated from the extracted SOM through micro-column chromatography, using activated silica gel. The aliphatic and aromatic fractions were then derivatized with 100 µL of N,O-bis (trimethylsilyl) trifluoracetamide and trimethylchlorosilane (BSTFA/TMS 99:1) at 65 °C for 30 min. Gas chromatography–mass spectrometry (GC–MS) analyses of the derivatized samples were performed on an Agilent model 6890 GC coupled to a Micromass GC-TOF MS (Chemistry Department at Iowa State University) and on an Agilent A 6890 N gas chromatograph/5973 network mass spectrometer (Department of Geological and Atmospheric Sciences at Iowa State University). Separation was achieved with a fused DB5 silica capillary column and with helium as the carrier gas. The GC operating temperature ramp was as follows: temperature was held at 65 °C for 2 min, and then increased from 65 to 300 °C at a rate of 6 °C/min, with final isothermal hold at 300 °C for 15 min. The sample was injected splitless with the injector temperature at 300 °C. The mass spectrometer was operated in the electron impact mode (EI) at 70 eV ionization energy and scanned from 40 to 650 Da. Individual compounds were identified by comparison of their mass spectra and retention times with those of published compounds and by interpreting mass fragmentation patterns. Relative abundances of the different compounds were calculated using
Some individual compounds were identified using the GC trace and the mass spectrum of a selected ion mass (SIM).

4. Results

4.1. Bulk geochemical parameters

4.1.1. Total organic carbon

Most of the samples analyzed are mudstones that vary in color from black or dark gray to slightly dark brown, typical for samples with relative high content of organic matter. Visible wood fragments were observed in some of the samples. Organic carbon contents (Corg) vary between 0.01 and 11.24 wt.% (Fig. 3). The lowest carbon contents were found in the Barco Formation (average Corg = 0.96 wt.%), corresponding to depths between 460 and 350 m below surface in the Site-A well. The highest contents were found in samples from the Cuervos Formation (average Corg = 2.00 wt.%), corresponding to depths between 350 and 150 m in the Site-A well.

4.1.2. Carbon isotopic composition of organic matter

The δ¹³C_TOM values for the Site-A samples range from −21.2‰ to −29.9‰, and they show three carbon-isotope excursions: a positive shift centered at 350 m and two negative shifts centered at 500 and 200 m (Fig. 4). The δ¹³C_TOM values for Core-B samples range from −24.3‰ to −270‰. The three carbon-isotope excursions found in the Site-A samples chronostratigraphically correlate with those found in marine carbonates (Zachos et al., 2001; Fig. 4). The δ¹³C_TOM values decrease from −24.2‰ to −26.5‰ from 65 to 58 Ma. This shift correlates with that in the δ¹³C_carbonate record from 0.5‰ to 2.5‰ for the same interval. A positive shift in δ¹³C_TOM values occurs in the Late Paleocene (58–56 Ma), with values becoming less negative from −26.5‰ to −23.8‰ in the δ¹³C_TOM and from 2.5‰ to 0.25‰ in δ¹³C_carbonate. One aspect that is absent in the δ¹³C_TOM record is the sharp spike in marine carbonate δ¹³C values at the Paleocene–Eocene boundary (~52 Ma), which corresponds to the PETM event (Koch et al., 1992; Zachos et al., 2001). Its absence in the δ¹³C_TOM data set is probably due to the large sampling intervals (~0.2 M.y./sample) employed for this study and the short duration (~0.02 to 0.07 M.y.) of this event (Smith et al., 2007).

4.2. Molecular composition of organic matter

Analyses of the aromatic fraction revealed undetectable concentrations of these compounds, possibly due to the high detection limit of the instrument utilized for the analysis of this fraction (Agilent A 6890 N gas chromatograph/5973 network mass spectrometer). Thus, this study reports the compounds identified in the aliphatic fraction of SOM. Four major families of organic compounds were identified in the aliphatic fraction: n-alkanes, regular acyclic isoprenoids, sesqui- and triterpenoids. Mid- and short-chain n-alkanes, and triterpenoids were the most abundant types of compounds present in the studied samples (Fig. 5).
4.2.1. n-Alkanes and isoprenoids

The n-alkane distribution in the studied samples shows an odd-over-even-predominance, with high abundances of short-chain (<nC20) n-alkanes (Fig. 6). The identified isoprenoids pristane (Pr) and phytane (Ph) are present in most of the samples, with Pr typically being more abundant than Ph. (e.g., Fig. 5). Because of their interdependent response to changes in the depositional environment, the Pr/Ph values (between 1.5 and 2.5) reflect Pr/Ph as a proxy for the level of oxicity in the sediments, with low interdependent response to changes in the depositional environment, being more abundant than Ph. (e.g., Fig. 5). Because of their interdependent response to changes in the depositional environment, the Pr/Ph values range from 0.94 to 4.59 with no visible trend with depth (Fig. 8). Site-A Pr/Ph values vary between 0.81 and 2.79 (Fig. 8), with a decreasing trend with depth (Fig. 8). Site-A Pr/Ph values range from 0.94 to 4.59 with no visible trend with depth (Fig. 9). These values suggest a changing level of oxygen in the sediments during the accumulation of the studied sequences.

4.2.2. Sesquiterpenoids and non-hopanoid triterpenoids

Two types of sesquiterpenoids were present in the saturated fraction of SOM: a C-16 sesquiterpenoid and a cadalene-type sesquiterpenoid. The identification of the two sesquiterpenoids was achieved by the presence of the characteristic fragments 183 and 213 in the mass spectra and by comparisons to published spectra of these compounds (Philp, 1985; Otto et al., 1997; Otto and Simoneit, 2001; Bechtel et al., 2003; Hautevelle et al., 2006). The non-hopanoid triterpenoids identified in the saturated fraction correspond to lupane-type triterpenoids.

4.2.3. Hopanoids

Hopanoids are compounds occurring in bacteria (Peters et al., 2005; Otto et al., 2005). These compounds, after n-alkanes, were the main constituents present in the saturated fraction of both Core-B and Site-A samples (Fig. 5). The identified hopanoids compounds were 17α-22,24,30-trisnorhopane, 17α(21H)-hopane, 17α(21H)-norhopane, 17β(21H)-hopane, 17α(21H)-homohopane, 17β(21H)-homohopane, unknown C32 hopanoid, and 17α(21H)-trishomohopane (H1-H2,H3,H4,H5,H6,H7,H8, respectively) (Fig. 5). Both set of samples display similar hopanoid distributions (Fig. 7), although the heavier hopanoids (H5 through H8) were commonly absent in the Core-B samples.

5. Discussion

5.1. The δ13CTOM values as a proxy for changes in the carbon cycle

The carbon isotopic composition of plant-derived organic matter (i.e., bulk organic matter, fossil wood, cuticles, etc.) has been used in geologic studies to evaluate the evolution of the carbon cycle through geologic times (e.g., Hasegawa, 1997; Gröcke et al., 1999; Hesselbo et al., 2003; Strauss and Peters-Kottig, 2003; Smith et al., 2007). The bulk organic matter in the Core-B and Site-A samples displays δ13C values around ~27‰ (Fig. 4), which are typical for C3 plants (Farquhar et al., 1989). The good correspondence between the marine and the terrestrial isotope data (Fig. 4) provides support to the notion that a tight linkage exists between the oceans, the atmosphere, and land plants, confirming that perturbations occurring between 50 and 65 Ma were global in extent. However, the implication of the parallelism that exists between the marine and terrestrial δ13C values still relies on the assumption that δ13CTOM values truly reflect δ13Cplant values. Although δ13CTOM values tend to reflect those of plant-derived organic matter (Lane et al., 2004; Rodelli and Gearing, 1984), other effects, including the extent of microbial alteration, could potentially alter δ13CTOM values by up to 2‰ (Gröcke, 1998, 2002; Bergen and Poole, 2002; Beerling and Royer, 2002). For instance, variations of ~1‰ between δ13CTOM and the isotopic composition of charcoal and lignite from the same stratigraphic horizons have been found (Heimböer et al., 2003; Gröcke et al., 2006). Moreover,
the sedimentological analysis of the studied deposits suggests that some sediments accumulated in environments where the contribution of aquatic plants to the sedimentary organic pool could be significant (e.g., oxbow lakes). The effect of diagenesis on carbon-isotopic values has been evaluated through different approaches, including those relying on comparisons between $\delta^{13}C_{\text{TOM}}$ and $\delta^{13}C$ values of plant cuticles (e.g., Arens and Jahren, 2000), $\delta^{13}C_{\text{TOM}}$ and $\delta^{13}C$ values of woody fragments (e.g., Gröcke, 2002) or on evaluations of the level of microbial degradation of fossil wood inferred from molecular components (biomarkers) (Bergen and Poole, 2002; Poole et al., 2004). For instance, Bergen and Poole (2002) identified high amounts of organic matter derived from bacteria as a result of the alteration of organic matter in fossilized woody fragments, which could potentially alter pristine $\delta^{13}C_{\text{plant}}$ values, thus constraining the use of $\delta^{13}C_{\text{TOM}}$ as a reliable proxy in estimating the evolution of the carbon cycle. To evaluate the role of diagenesis and the sources of organic matter on $\delta^{13}C_{\text{TOM}}$ values, several biomarker ratios were employed in this study because of their demonstrated response to diagenesis and/or their

![Fig. 7. Histograms of the distribution patterns of hopanoids in representative samples of Core-B (lower panel, depth = 509.9 m) and Site-A (upper panel, depth = 548.6 m). The identified hopanoids compounds correspond to 17α-22,24,30-trinorhopane, 17α(21H)-hopane, 17α,21(21H)-norhopane, 17α(21H)-hopane, 17α(21H)-homohopane, 17α(21H)-homohopane, unknown C32 hopanoid, and 17α(21H)-trishomohopane (H1,H2,H3,H4,H5,H6,H7,H8, respectively).](image)

![Fig. 8. Stratigraphic variability of the Carbon Preference Index (CPI), pristane/phytane (Pr/Ph) ratio, aquatic/terrestrial plants (Paq) ratio, and smoothed carbon isotopic composition ($\delta^{13}C$) in the Core-B.](image)
specifity to different sources of organic matter (e.g., vascular plants, algae, bacteria). These ratios were calculated from the relative abundances of the different compounds identified in the saturated fraction of the SOM. The saturated fraction shows that n-alkanes display an overall odd-over-even predominance (Figs. 8 and 9), which is usually associated with significant input of organic matter from terrestrial vascular plants (Bechtel et al., 2003; Otto et al., 2005). Odd number long-chain n-alkanes are major components of plant cuticular waxes formed as a result of elongation and further decarboxylation from a fatty acid precursor (e.g., palmitate) (Harwood and Russel, 1984). Although this odd-over-even predominance of n-alkanes is a good indicator of terrestrial contributions, Peters et al. (2005) suggested that a better evaluation of the potential contribution of land plants to the bulk organic matter can be achieved with the Carbon Preference Index (CPI) ratio, which is determined through the following equation:

$$\text{CPI} = \frac{1}{2} \left( \frac{25 + 27 + 29 + 31 + 33}{24 + 26 + 28 + 30 + 32} \right) + \frac{1}{2} \left( \frac{25 + 27 + 29 + 31 + 33}{26 + 28 + 30 + 32 + 34} \right).$$

The numbers in the equation represent the number of carbons in an n-alkane molecule. CPI values for the Site-A and the Core-B samples range from 1.24 to 1.92 and from 1.6 to 2.47, respectively (Figs. 8 and 9). While CPI values of the Site-A samples increase with depth (Fig. 9), those of the Core-B samples show no trend and they fall within a narrow range (1.7–2.1), with the exception of the values at 470 m (Fig. 8). These obtained CPI values are lower than those commonly observed for extant vascular plants (Van Dongen et al., 2006), which are commonly >3, but they do suggest significant contributions of organic matter derived from these higher plants, since CPI values greater than 1 correspond to a predominantly land-plant input (e.g., Ficken et al., 2000; Schefuš et al., 2003; Muri et al., 2004; Peters et al., 2005; Van Dongen, 2006). Although both n-alkane distribution and CPI values suggest the predominance of vascular plant-derived organic matter, the presence of short chain lipids in the saturated fraction in significant abundances (~80% in average, Figs. 6 and 7) suggests other type of contributions different than those of terrestrial land plants. Short-chain lipids are commonly associated with the input of organic matter derived from freshwater photosynthetic algae and/or macrophytes (submerged/floaters or macrophytes) (Cranwell et al., 1987; Mello and Maxwell, 1990; Bechtel et al., 2003; Muri et al., 2004; Van Dongen et al., 2006). Because some aerobic bacteria decompose organic matter when anoxic conditions are not rapidly reached, short-chain lipids could also come from such organisms (e.g., Cranwell et al., 1987; Bechtel et al., 2003), and their contribution to the studied sediments cannot be ruled out.

Despite the potential influence of lacustrine algae and/or bacterial organic matter, the odd-over-even predominance showed by the CPI index shows that, within the high molecular weight n-alkanes, landplant-derived organic matter contributed significantly to the bulk of the straight-chain compounds. The compounds nC25, nC27, nC29, and nC31 are the dominant compounds, rather than even number long-chain n-alkanes, which are synthesized by bacteria and marine organisms (Tissot and Welte, 1984; Peters et al., 2005).

Because of the abundance of short-chain lipids and their possible origin (freshwater photosynthetic organisms), the terrestrial/freshwater plants (Paq) ratio developed by Ficken et al. (2000) can be used to assess the source of most of the long-chain lipids preserved in the studied sediments. The Paq ratio is defined as the ratio between the abundance of mid-chain n-alkanes (nC23, nC25) produced by submerged/floaters (macrophytes) over the amount of long-chain n-alkanes (nC27, nC29, nC31) produced by terrestrial plants:

$$P_{aq} = \frac{(23 + 25)}{(27 + 29 + 31)}.$$

Paq values <0.4 suggest a predominant terrestrial input, values >0.75 reflect a primary submerged/floaters plants contribution, and values between 0.4 and 0.75 reflect a mixture. The overall decreasing trend in Paq values towards shallower (younger) depths indicates a transition towards more terrestrial conditions in younger samples (Figs. 8 and 9). Core-B samples display lower values relative to those from the Site-A 1 samples, suggesting a higher input of terrestrial contributions.

Fig. 9. Stratigraphic variability of the Carbon Preference Index (CPI), pristane/phytane (Pr/Ph) ratio, aquatic/terrestrial plants (Paq) ratio, and smoothed carbon isotopic composition (δ13C) in the Site-A well.
plants to the bulk organic matter into the Cesar–Rancheria basin or the result of a poor preservation conditions (i.e., conditions not allowing plant-derived organic matter to be preserved in sediments) in the Catatumbo basin (Site-A). This index, thus, suggests a mixed contribution of terrestrial and freshwater plants.

Supporting the conclusion that higher plant-derived organic matter is present in the studied sediments, sesquiterpenoids were identified in the saturated fraction (Fig. 5). Bicyclic sesquiterpenoids have been identified in a variety of geological materials, from recent and ancient sediments to coals, oils, peats, ambers, and fossil resins (Otto et al., 1997; Otto and Simoneit, 2001; Bechtel et al., 2002, 2003; Tuo and Philp, 2005; Hautevelle et al., 2006). Sesquiterpenoids are widely distributed among vascular plants, including both angiosperms and gymnosperms (Otto and Simoneit, 2001). The exception is cadalene, which is a compound that appears to be related to gymnosperm-derived material (Otto et al., 1997; Bechtel et al., 2003). Although cadalene has been recognized as a major component of resins in several conifer (gymnosperm) species (Philp, 1985; Otto et al., 1997), it has also been reported to result from the degradation of resins produced by some angiosperm species (Otto et al., 1997).

Both cadalene and the C_{16} sesquiterpane were identified in the studied samples (Tp1 and Tp2, Fig. 5), suggesting the contribution of gymnosperm-derived organic matter to the studied sediments. In addition to the possible contribution of gymnosperms, the following angiosperm-associated triterpenoids were found in the saturated fraction: lupane- (L_{2}) and normoretane-type (L_{1}) compounds. These two compounds have been found in leaf, wood, root, and bark tissues of these plants (Sukh, 1989; Bechtel et al., 2003). The presence of these compounds in the studied sediments, coupled with the presence of sesquiterpenoids, supports the CPI data, suggesting that vascular plants were an important source of the organic matter in the Site-A and Core-B sediments.

5.2. Preservation of the organic matter

Although CPI values reflect a slightly dominant contribution from terrestrial plants to the bulk organic matter, the predominance of short-chain over long-chain n-alkanes could also result from a poor preservation of the heavier compounds. Redox conditions govern to a large extent the preservation potential of organic matter, with oxic conditions leading to poor preservation. Because of its sensitivity to redox conditions, the Pr/Ph ratio can be used to evaluate the effect of oxicity during the accumulation of the studied samples (Didyk et al., 1978; Bechtel et al., 2003).

Variations in the Pr/Ph ratios for the studied samples (Figs. 8 and 9) suggest the existence of two different redox regimes governing the depositional settings of the two basins (Catatumbo basin for Site-A and Cesar–Rancheria basin for Core-B), Core-B samples, covering the time...
interval from ~66 to 57 Ma, show Pr/Ph values indicative of a change from anoxic (below 450 m) to dysoxic (450–350 m) and possibly oxic conditions (above 350 m). Site-A Pr/Ph values suggest that the organic matter in those sediments was deposited under dysoxic to oxic conditions between 450 and 600 m (Catatumbo Formation), changing to anoxic/dysoxic conditions between 450 and 300 m (Barco Formation), and shifting towards more oxic conditions at depths above 300 m of the section (Cuervos Formation).

This increase in oxic conditions leads to the degradation of less resistant compounds by microorganisms, which can be evaluated through the compounds that they produce (Peters et al., 2005). Hopanes (H1–H8, Figs. 5 and 7) are important constituents of the saturated fraction in the studied samples, and these compounds are associated with microbial contributions to the bulk organic matter (Peters et al., 2005). For that reason, hopane abundances can be used to estimate the intensity of biomass degradation by using the ratio $\beta_3/(\beta_3 + \alpha_3)$ hopanes (Fig. 10) (MacKenzie et al., 1981; Bechtel et al., 2003; Van Dongen et al., 2006).

Commonly, low $\beta_3/(\beta_3 + \alpha_3)$ values (below 0.5) are indicators of moderate to high degradation of organic matter, because $\alpha_3$ hopanes are more kinetically stable as diagenesis degrades sedimentary organic matter. Fig. 10 shows that Site-A and Core-B samples display parallel trends, which suggest that, despite the differences in preservation and contribution from terrestrial sources, the organic matter at both sites experienced similar degradation patterns.

![Fig. 12](image_url)

**Fig. 12.** Cross correlation of non-smoothed $\delta^{13}C_{\text{OM}}$ vs. Pr/Ph, $\beta_3/(\beta_3 + \alpha_3)$, CPI, and $P_{aq}$ values for the studied sections. Site-A (upper panel) relationships for $\delta^{13}C_{\text{OM}}$ vs. Pr/Ph, $\beta_3/(\beta_3 + \alpha_3)$, CPI, and $P_{aq}$ are not significant ($p < 0.87, 0.7446, 0.81$, and $0.77$, respectively). Core-B (lower panel) relationships for $\delta^{13}C_{\text{OM}}$ vs. Pr/Ph, $\beta_3/(\beta_3 + \alpha_3)$, CPI, and $P_{aq}$ are not significant ($p < 0.96, 0.9762, 0.54$, and $0.076$, respectively).
5.3. Evaluation of $\delta^{13}$CTOM values

The CPI values and the sesquiterpenoid and lupanoid abundances confirm the assumption of a predominant terrestrial origin for the bulk organic matter in the studied sediments, which is also supported by the pollen data. However, the predominant dysoxic–oxic conditions during deposition (as determined from the Pr/Ph values) and the significant levels of biomass degradation during diagenesis (as interpreted from $[\beta/\beta']$ and $[(\beta'+\gamma)/\alpha\beta']$) values) could imply that the measured $\delta^{13}$CTOM values are different than the original $\delta^{13}$CTOM values. To evaluate this potential effect, $\delta^{13}$CTOM values were plotted against our diageneric proxies (i.e., Pr/Ph, Cpap, and $[\beta/\beta'+\gamma]$). Additionally, $\delta^{13}$CTOM values were also plotted against CPI and Pch to evaluate the possible influence of the type of organic matter on the observed trend in $\delta^{13}$CTOM values.

No significant correlation exists between each of the parameters analyzed and $\delta^{13}$CTOM values (Figs. 11 and 12), indicating that neither depositional environment nor the degree of biomass alteration during diagenesis has significantly altered the measured $\delta^{13}$CTOM values. Consequently, considering that similar results were obtained from the organic matter of two different basins, this study indicates that the measured $\delta^{13}$CTOM values could be used for chronostratigraphic purposes, since they are possibly close to those of the ancient plants. The higher CPI could account for differences in the amount of land-plant contribution to pollen data. However, the predominant dysoxic–oxic organic matter in the studied sediments, which is also supported by the paleoenvironmental tool. Lethaia 31, 1–18.

Appendix A. Supplementary data

Supplementary data associated with this article can be found in the online version, at doi:10.1016/j.palaeo.2009.03.015.

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