Recognizing the Albian-Cenomanian (OAE1d) sequence boundary using plant carbon isotopes: Dakota Formation, Western Interior Basin, USA

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ABSTRACT

Analysis of bulk sedimentary organic matter and charcoal from an Albian-Cenomanian fluvial-estuarine succession (Dakota Formation) at Rose Creek Pit (RCP), Nebraska, reveals a negative excursion of ~3‰ in late Albian strata. Overlying Cenomanian strata have $\delta^{13}C$ values of −24‰ to −23‰ that are similar to pre-excursion values. The absence of an intervening positive excursion (as exists in marine records of the Albian-Cenomanian boundary) likely results from a depositional hiatus. The corresponding positive $\delta^{13}C$ event and proposed depositional hiatus are concordant with a regionally identified sequence boundary in the Dakota Formation (D_2), as well as a major regressive phase throughout the globe at the Albian-Cenomanian boundary. Data from RCP confirm suggestions that some positive carbon-isotope excursions in the geologic record are coincident with regressive sea-level phases. We estimate using isotopic correlation that the D_2 sequence boundary at RCP was on the order of 0.5 m.y. in duration. Therefore, interpretations of isotopic events and associated environmental phenomena, such as oceanic anoxic events, in the shallow-marine and terrestrial record may be influenced by stratigraphic incompleteness. Further investigation of terrestrial $\delta^{13}C$ records may be useful in recognizing and constraining sea-level changes in the geologic record.

Keywords: oceanic anoxic event, carbon isotopes, plants, sequence boundary, Albian-Cenomanian, middle Cretaceous.

INTRODUCTION

Carbon-isotope ratios ($\delta^{13}C$) of bulk terrestrial organic matter and isolated individual plant fragments have recently been used to reconstruct terrestrial plant ecologies (Nguyen Tu et al., 1999), interpret isotopic changes in atmospheric CO_2 (Gröcke et al., 1999), estimate atmospheric pCO_2 changes (Hasegawa et al., 2003), and assign terrestrial sequences with marine biostratigraphies (Hesselbo et al. 2000). Isotopic investigations of Cretaceous oceanic anoxic events (OAEs) indicate significant shifts in the global carbon cycle, most notably during the early Aptian (OAE1a) and at the Cenomanian-Turonian boundary (OAE2). OAEs are characterized by relatively long-duration positive $\delta^{13}C$ excursions (Arthur et al., 1985) that are preceded by rapid negative $\delta^{13}C$ excursions (Leckie et al., 2002), with the exception of OAE2. OAEs with negative $\delta^{13}C$ excursions have been documented throughout the Cretaceous (Herrle et al., 2003), but it is not yet known whether these events are global in distribution and affect, although negative $\delta^{13}C$ excursions associated with OAE1a and OAE1b (Aptian-Albian boundary) are recorded in the terrestrial carbon reservoir (Gröcke et al., 1999; Heimhofer et al., 2003).

To ascertain the effect of OAE1d (latest Albian) on the terrestrial carbon-isotope record, we collected bulk sediment and plant macrofossils for organic carbon-isotope analysis from the Rose Creek Pit (RCP) in the Dakota Formation, near Fairbury, Nebraska (Fig. 1). Palynostratigraphy places the section at RCP as late Albian to Cenomanian in age (Brenner et al., 2000). A direct comparison between our terrestrial $\delta^{13}C$ records from RCP and the oceanic $\delta^{13}C_{foram}$ record at ODP Site 1052 (Wilson and Norris, 2001) provides better time reso-

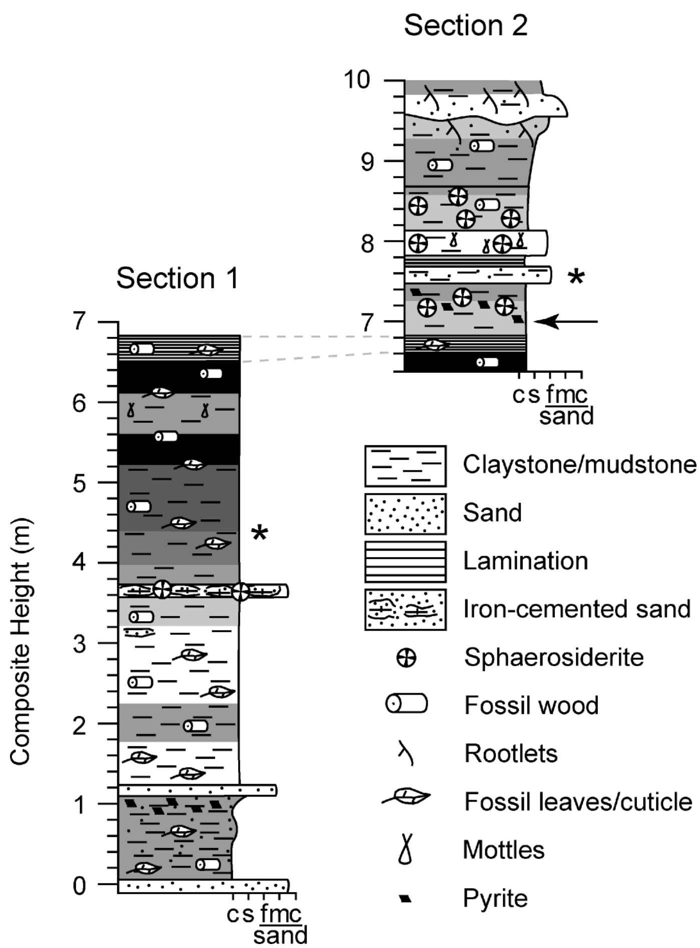

Figure 1. Two overlapping stratigraphic sections were combined in this study of Rose Creek Pit (RCP), Nebraska. Stars represent sample positions for palynological analysis. Arrow points to the D_2 sequence boundary identified by this study in carbon isotopes; c—clay, s—silt, fmc—fine, medium, and coarse sand.

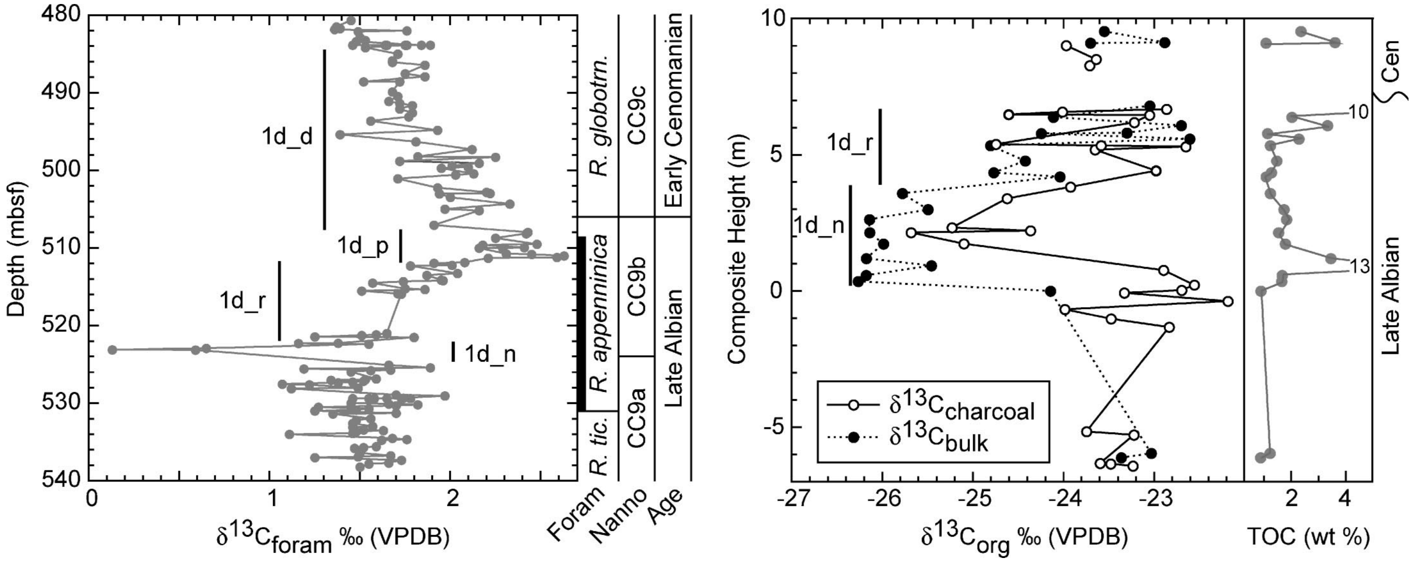

Figure 2. A: $\delta^{13}C_{foram}$ curve from ODP Site 1052, Blake Nose (Wilson and Norris, 2001), showing a rapid negative excursion in the late Albian followed by a positive excursion that peaks just before Albian-Cenomanian boundary. Labeled excursion segments are 1d_n—negative; 1d_r—rise; 1d_p—positive; and 1d_d—decline. Foraminifera and nannofossil biostratigraphy is according to Wilson and Norris (2001). Solid black bar in *R. appenninica* zone gives the occurrence of laminated black shales. Note that initiation of the OAE1d negative $\delta^{13}C$ excursion occurs at boundary between nannofossil zones CC9a and CC9b. B: $\delta^{13}C_{bulk}$ and $\delta^{13}C_{charcoal}$ curves from Rose Creek Pit, Nebraska. Palynology is determined from bulk samples at 4.5 and 7.6 m. TOC values were determined during isotopic analysis of decarbonated bulk sediment and show two phases of organic richness, beginning at onset of the negative $\delta^{13}C$ excursion and just prior to the isotopically identified hiatus. Stratigraphic samples below 0 m were obtained from a nearby section on highway at RCP: This interval was dominated by sand and thus few charcoal samples were obtainable.

lution within the Dakota Formation (Figs. 2A and 2B). Our data indicate that the regressive phase associated with OAE1d occurred over a period of <0.5 m.y. and was a potential driver for paleoceanographic phenomena (Watkins et al., 2005), such as disruption of a stratified ocean and an extinction event. This result allows us to assess the impact in the ocean-atmosphere system associated with OAE1d.

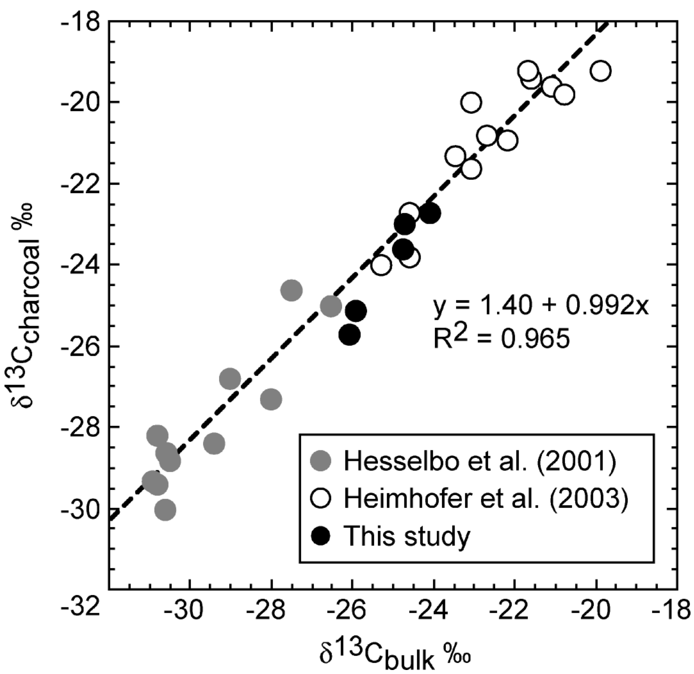

Figure 3. $\delta^{13}C$ of charcoal fragments versus bulk sediment for Aptian-Albian from Algarve Basin, Portugal (Heimhofer et al., 2003), and Rose Creek Pit, Nebraska (this study). For comparison, wood and bulk sediment samples from the Early Toarcian oceanic anoxic event (Jurassic) have been included (Hesselbo et al., 2000). Each point is representative of samples from same stratigraphic horizon.

GEOLOGICAL SETTING AND METHODOLOGY

The RCP locality has previously been described by Basinger and Dilcher (1984) and Upchurch and Dilcher (1990). On the basis of palynoflora at RCP, Farley and Dilcher (1986) concluded that the deposit is Cenomanian in age. Continuing studies, however, have shown that the RCP locality straddles the Albian-Cenomanian stage boundary and contains the D_2 sequence boundary of Brenner et al. (2000). The D_2 surface unconformably separates deposits of the underlying Upper Albian Muddy Cycle (''J'' Sandstone) from the overlying Cenomanian-Turonian Greenhorn Cycle (''D'' Sandstone) (Weimer, 1987). Our palynologic samples from 4.5 m at RCP (Fig. 1) produce definitive Upper Albian spores *Disaltriangulisporites perplexus* and *Podocarpidites multesimus*, along with numerous probable, but not necessarily definitive Albian palynomorphs. Significantly, no Cenomanian taxa are recognized in our palynologic analysis of the leaf beds. Samples from 7.6 m at RCP (Fig. 1), however, produce definitive Lower Cenomanian forms *Foveogleicheniidites confossus* and *Artiopollis indivisus*, thereby bracketing the position of the D_2 sequence boundary.

Bulk sediment and macro-charcoal samples were collected throughout the 10 m section at RCP. Both bulk-sediment and charcoal samples were prepared according to the methods outlined in Gröcke et al. (1999). Carbon-isotope values are reported in the standard delta (δ) notation relative to VPDB with analytical precision better than 0.1‰. Micromorphologic investigations of pedogenically modified mudstones at the Albian-Cenomanian sequence boundary were carried out on thin sections cut from resin-impregnated billets, using methods and terminology from Bullock et al. (1985).

TERRESTRIAL-MARINE CORRELATION

An Albian-Cenomanian foraminiferal oceanic carbon-isotope curve from ODP Site 1052 shows a pronounced negative excursion at the boundary between nannofossil subzones CC9a and CC9b, ~100 k.y. after the depositional onset of laminated black shales (Fig. 2A) (Wilson and Norris, 2001). In order to use isotope stratigraphy as a correlative tool, we define and label segments of the $\delta^{13}C_{foram}$ curve that describe its structure. The OAE1d $\delta^{13}C_{foram}$ curve shows a rapid,

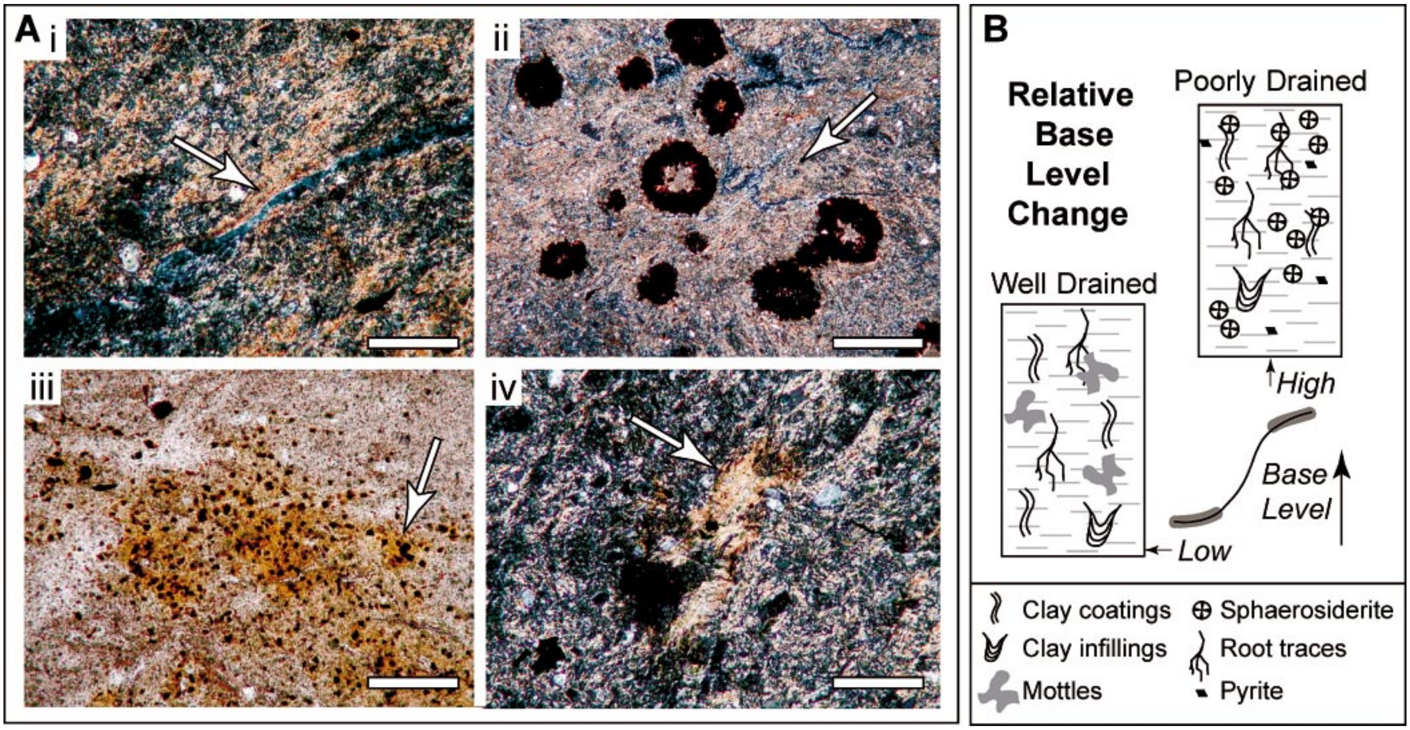

Figure 4. A: Photomicrographs of paleosol at D_2 sequence boundary. Scale bars = 0.2 mm, except for where indicated. (i) Cross-polarized light image of claystone paleosol with striated birefringent fabric. Bright birefringent streaks consist of oriented illuvial clay accumulations reworked from void coatings into matrix. Arrow indicates argillan (pedogenic clay coating). (ii) Cross-polarized light image of striated birefringent fabric (arrow) with overprinting sphaerosiderite nodules. Scale bar = 1 mm. (iii) Plane-polarized light image of 0.01–0.02 mm pyrite framboids (arrow) within fine, faint yellow mottled domains of claystone. (iv) Cross-polarized light image of degraded clay infilling (arrow) with wavy to grainy extinction patterns and splayed, smeared boundaries. B: Schematic illustration shows how drainage conditions changed within the paleosol as base level changed during development of D_2 surface. In general, pedogenesis was influenced by an early, stable phase characterized by episodic wetting and drying, a fluctuating water table, and relatively low base level. Later, base level was high, resulting in saturated (reducing redox) conditions, minor pyrite precipitation (perhaps due to marine- and meteoric-phreatic water mixing), and sphaerosiderite precipitation.

pronounced negative excursion (Fig. 2A: 1d_n), followed by a rapid rise (Fig. 2A: 1d_r) to the most positive values (Fig. 2A: 1d_p) that peak just prior to the Albian-Cenomanian boundary, after which $\delta^{13}C_{foram}$ values gradually decline (Fig. 2A: 1d_d) to preexcursion values.

The RCP $\delta^{13}C$ curve for bulk organic matter and charcoal record a large rapid negative excursion of ~3‰ in late Albian strata (Fig. 2B). This is followed by a gradual return to preexcursion $\delta^{13}C$ values and the deposition of a plant-rich lignitic mud. Comparison of the terrestrial and oceanic records show that the major positive $\delta^{13}C$ excursion at the Albian-Cenomanian transition is missing in the strata at the RCP locality, implying the existence of a hiatus. Lower Cenomanian $\delta^{13}C$ values are similar to preexcursion values (−24‰ to −23‰). The bulk organic matter $\delta^{13}C$ record is, in most cases, more negative than charcoal $\delta^{13}C$ values (Fig. 2B: 1d_n) and potentially relates to compositional differences in the bulk rock. However, during the deposition of the lignitic mud, the $\delta^{13}C$ values for both bulk organic matter and charcoal shift rapidly, perhaps in relation to a change in organic matter composition.

In the Atlantic and Tethys Oceans, organic-rich black shales initiated in the marine realm ~100 k.y. prior to the onset of the rapid negative excursion (Fig. 2A). If OAEs are to be associated with ''global productivity events,'' then a similar stratigraphic deposit (under the right circumstances) should occur in, at least, some terrestrial sequences. A TOC record of bulk sedimentary organic matter from RCP was produced in order to assess this. Two phases of elevated TOC levels were recorded, with the first coinciding with the onset of the negative $\delta^{13}C$ excursion. Although it is tempting to suggest that a possible global carbon burial event occurred in both the marine and terrestrial realms, this study focuses on one site that is dominated by ponding, and thus the TOC record may be controlled by local processes only.

BULK VERSUS CHARCOAL $\delta^{13}C$ RECORDS

Heimhofer et al. (2003) found that leaf and cuticle $\delta^{13}C$ values were similar to bulk organic-matter values from an Aptian-Albian shallow-marine sequence in Portugal. That data set recorded an offset of +1.7‰ between co-occurring samples of charcoal/lignite and bulk organic matter. Through the RCP section, an average offset of +1.4‰ is recorded. Figure 3 illustrates this offset, including data from the Early Toarcian (Jurassic) oceanic anoxic event (Hesselbo et al., 2000). This offset is striking ($R^2 = 0.965$) in that it is robust and consistent irrespective of sedimentary facies, interpreted climate, environment, age, and diagenetic history.

Heimhofer et al. (2003) attributed the isotopic difference to bulk sediment consisting predominantly of cuticle and leaves. The sediments at RCP are very rich in leaf and cuticle fragments (Fig. 1), and it is known that the difference between leaves and wood (a dominant source of charcoal) is ~+3‰ (Leavitt and Long, 1991). Since the average difference between $\delta^{13}C_{bulk}$ and $\delta^{13}C_{charcoal}$ is consistent (Fig. 3), other mechanisms must also be contributing to the +1.4‰ shift. Because the offset is similar in three very different data sets (this study, Hesselbo et al. [2000], and Heimhofer et al. [2003]), an environmental parameter (such as local hydrology or climate) must be ruled out. In charcoalification experiments, $\delta^{13}C_{charcoal}$ values become more ^{12}C-enriched (~1.2‰) than the original material (Jones and Chaloner, 1991). Hence, these two factors (leaves versus wood and charcoalification) would suggest an average shift of ~+1.8‰, which is similar to the data set presented in Figure 3. More data and multicomponent analysis of plant materials from stratigraphic horizons are required to fully ascertain the cause behind this relationship in the geologic record.

SEA LEVEL, D_2 SEQUENCE BOUNDARY, AND $\delta^{13}C$

Gröcke et al. (1999) noted that the Early Aptian positive $\delta^{13}C$ excursion corresponded with a relative sea-level fall in the tectonically active Wessex Basin, Isle of Wight, and in other sections around the world. The isotopic study of the RCP section presents a similar scenario: The positive excursion is absent (Fig. 2B) because of a regional/global regressive event (Weimer, 1987). If positive $\delta^{13}C$ excursions are associated with regressive phases, then the terrestrial and shallow-marine environmental record may be skewed and missing as illustrated in this study. This consideration is critical because the majority of terrestrial-isotope records have been constructed from shallow-marine successions that are more likely to be influenced by sea-level changes (e.g., Hasegawa et al., 2003; Heimhofer et al., 2003).

The RCP sample interval from 6.6 to 8.6 m (Fig. 1), at the position of the interpreted D_2 sequence boundary (Fig. 2B), consists of a gleyed mudstone paleosol with block structure, slickensides, and abundant sphaerosiderite. Micromorphologic studies of this interval (Fig. 4A) show evidence for clay illuviation during an earlier, more well-drained phase of paleosol development. This interval was overprinted by later hydromorphic conditions, under which sphaerosiderite and pyrite framboids were precipitated. Macroscopic blocky soil structure is greater than that seen above and below this interval. Subsequent oxidation of the reduced iron minerals appears to be the result of modern weathering. The overprinting of early vadose textural pedofeatures, such as clay cutans and microlaminated clay infills (Fig. 4B) by later hydromorphism, suggests that the D_2 exposure surface at RCP experienced a polygenetic history of soil development (McCarthy and Plint, 1998). White et al. (2005) showed that changes in the stable isotopic and major element chemistry of sphaerosiderites at the D_2 surface in the Dakota Formation of Iowa are indicative of major paleohydrologic changes in the coastal plain as a result of base-level change (Fig. 4B), an idea that could be tested further at RCP.

The D_2 sequence boundary at the Albian-Cenomanian boundary is recognized around the globe (Scott et al., 2000; Bachmann et al., 2003), but the duration of this event is uncertain. The isotopic record at RCP suggests that the regressive event initiated during the latest Albian (equivalent to upper *R. appenninica* foraminifera and CC9b nannofossil zones). Using the time scale derived in Wilson and Norris (2001) and the trend of the isotopic curves (Figs. 2A and 2B), the minimum duration of this sequence boundary at RCP would be on the order of $\sim$0.5 m.y. The short-lived consequence of this regressive phase coincides with significant marine environmental phenomena, such as a breakdown in oceanic stratification and an extinction event (Watkins et al., 2005). Glacio-eustatic causes have been proposed as a cause for rapid sea-level changes. Although we do not have direct evidence for such a phenomenon as a causal mechanism of the D_2 sequence boundary, recent studies on stratigraphic successions in India suggest the possibility of Albian-Cenomanian glacial cycles (Gale et al., 2002).

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