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ORIGINAL ARTICLE

Carbon isotope excursion at Paleocene–Eocene transition in Jaisalmer Basin, western Rajasthan, India

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Abstract A negative carbon isotope excursion at Paleocene–Eocene boundary in Jaisalmer Basin has been recognized for the first time. Besides excursion at P–E transition, short but significant negative carbon isotope shift has also been observed preceding the Paleocene–Eocene thermal maximum (PETM). The study discusses the possible mechanism for PETM and pre-CIE warm conditions in Jaisalmer Basin, India.

Keywords Carbon isotope · Jaisalmer Basin · Paleocene–Eocene boundary · Excursion

Introduction

Close to the boundary between the Paleocene and Eocene epochs, approximately 55.5 Ma ago (Berggren et al. 1992), a distinct phase of global warming occurred, which has been called the Paleocene–Eocene thermal maximum (PETM), and which was superimposed on already warm conditions. Associated with the warming is a negative 2.5–6‰ carbon isotope ($\delta^{13}C$) excursion (CIE) (Kennett and Stott 1991; Koch et al. 1992; Pagani et al. 2006; Thomas et al. 2002), generally accepted to reflect the geologically rapid injection of C-depleted carbon, in the form of CO and/or CH into the global exogenic carbon pool. The duration of the PETM, as defined by the negative carbon isotope excursion and subsequent recovery, is still debated (Rohl et al. 2000; Bowen et al. 2001; Farley and Eltgroth 2003). Substantial information is available on P–E transition from Northern Hemisphere localities. Recognition of equivalent successions in the Southern Hemisphere is rather sparse, maybe due to limited land mass distribution. C and O isotope results of subsurface samples from Chinnawala well (CT-1, ONGC; Fig. 1) from Jaisalmer Basin, western Rajasthan, India are reported here.

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Geology

The Rajasthan Shelf is located to the west of the Aravalli ranges and forms the eastern flank of the Indus Shelf. The sedimentary tract to the west and northwest of Aravallis up to Pakistan border is divided into three basins (Fig. 1): Jaisalmer Basin, Bikaner-Nagaur Basin and Barmer Basin. The Precambrian rocks form the floor of the sedimentary successions deposited in the western Rajasthan basins. The Jaisalmer Basin has both Mesozoic and Cenozoic sediments exposed along the eastern margin. It is a pericratonic basin bounded toward east and southeast by Barmer–Devikot–Nachna High and at the southern limit by Barmer High. Various organizations have undertaken drilling in this basin for exploration of hydrocarbons and in turn could provide subsurface samples and data to understand the stratigraphy of this basin more precisely. Misra et al.

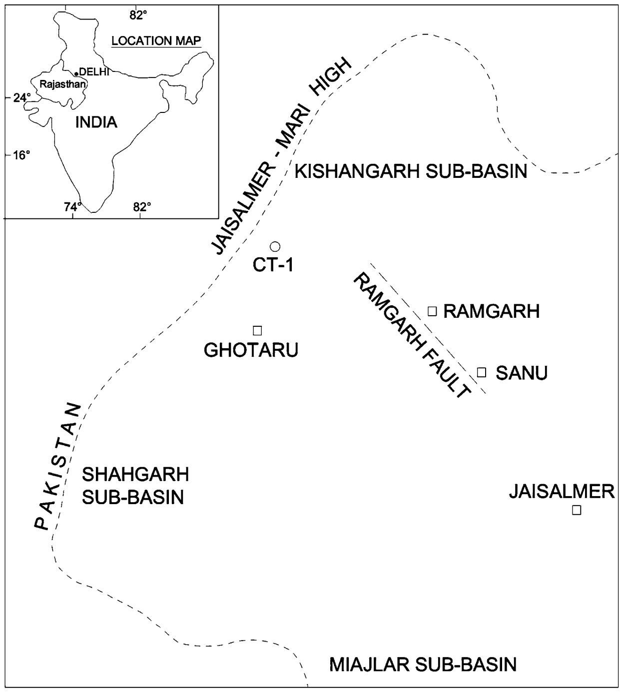

Fig. 1 Location map of the study area, western India

(1993) have established 16 litho units of formation rank in the Phanerozoic succession of Jaisalmer Basin.

Figure 2 presents the lithological, chronological and inferred geological events of the Sanu and Khuiala Formation of Jaisalmer Basin. The Sanu Formation in subsurface is divisible into two members (Dasgupta et al. 1973): The lower Mohmad Dhani member rests unconformably over Mesozoic, where it comprises a thick sequence of sandstone. The upper Kharatar member comprises marl, shale and limestone, which are absent in the outcrop. The Khartar member is rich in planktons, larger foraminifera and ostracodes, which suggest a Late Paleocene age. The lower member was deposited in the Continental environment, while the upper member was laid during the inner to middle Neritic. The Khuiala Formation overlies Sanu Formation with disconformable contact in exposed sections, but is conformable in the subsurface (Narayanan 1959). The formation comprises limestone and shale interbedded with thin argillaceous limestone. The formation is rich in foraminifera and ostracodes, suggesting an age of Late Paleocene to Early Eocene. An inner to middle Neritic environment has been proposed for the Khuiala Formation.

Chidambram 1991; Misra et al. 1996; Bhandari 1995, 1996; Pandey and Dave 1998 and many other researchers have contributed significantly to erect the geological history of Jaisalmer Bain. Since the Early Cambrian, frequent alternative transgression, regression and hiatus periods were continued until the Late Eocene. Marine water completely withdrew from this basin, and a major hiatus spanning from Late Eocene to Pleistocene to Recent is recognized. A major transgression was observed during deposition of Upper Goru (late Albian–Cenomanian) with a maximum bathymetry of about 100–150 m. A drop in sea level was observed near the top of upper Goru (Cenomanian). The overlying Parh Formation (Turonian–Coniaacian) was laid under a major transgressive phase. A major uplift in the Indus Basin resulted in a withdrawal of the sea, and a subsequent major hiatus spanning from Santonian (Late Cretaceous) to Danian (Early Paleocene) marked the end of Mesozoic maine sedimentary cycle. The Tertiary sedimentation began with the deposition of a clastic sequence, the Sanu Formation (Danian). These sediments were laid under continental to shallow marine environment and developed good reservoir facies. This in turn is overlain by transgressive phase represented by the Kharatar Member. This is followed by regression near the top of Late Paleocene. A rise in sea level was observed during deposition of the Khuiala Formation.

Analytical methods and evaluating data quality

Subsurface samples were considered. Least altered samples (lacking veins, discoloration, weathered rinds, recrystallization features and silicification) were microdrilled with a 1-mm drill. CO_2 was extracted from these carbonate samples in a high vacuum line after reaction with phosphoric acid at 25°C and cryogenically cleaned according to the method described by Craig (1957). CO_2 gas released was analyzed for O and C isotopes in a double inlet, triple collector mass spectrometer (VG Isotech SIRA II), using the BSC reference gas (Borborema skarn calcite) that calibrated against NBS-18, NBS-19 and NBS-20 and having a δ^{18}O value of $-11.3‰_{PDB}$ and $\delta^{13}C = -8.6‰_{PDB}$. The external precision based on multiple standard measurements of NBS-19 was better than 0.1‰ for carbon and oxygen. The results are expressed in the δ-notation in parts per thousand in relation to international PDB scale.

C-isotope compositions of most of the bulk rock probably did not change much during diagenesis, because the volume of carbon within the carbonate rock was vastly greater than that in the pore-water reservoir and because the fractionation between calcium carbonate was relatively small at near-surface temperatures (Emrich et al. 1970; Scholle and Arthur 1980). Quite a few parameters have been proposed to evaluate the degree of alteration of individual samples (Kah et al. 1999 and Kaufman and Knoll 1995; Melezhik et al. 2001). Oxygen isotopes are sensitive indicators of alteration, with marked decrease in δ^{18}O values if the system undergoes isotopic exchange with meteoric or hydrothermal fluids. In diagenetic

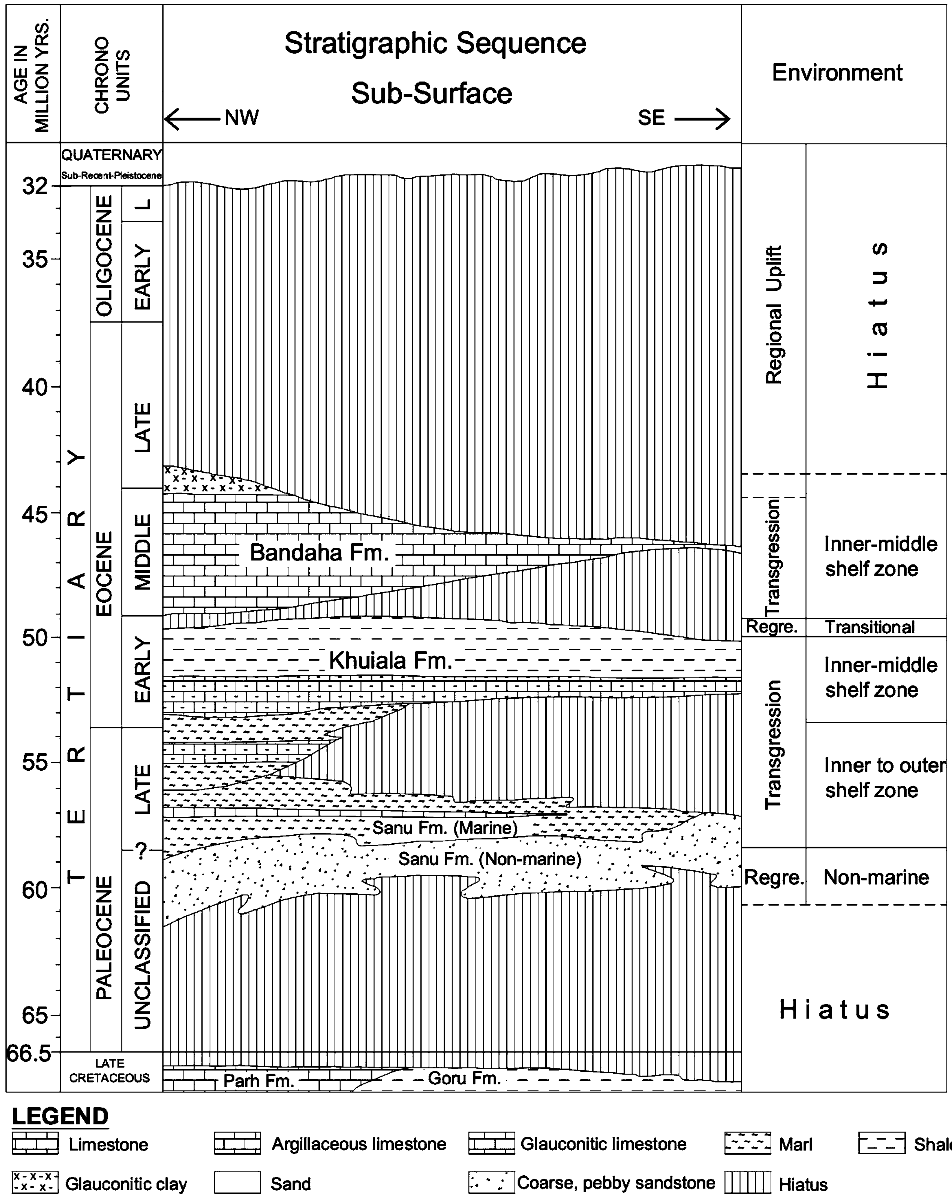

Fig. 2 Subsurface Tertiary stratigraphy of Jaisalmer Basin, India

environments, three orders of magnitude higher water–rock ratios are supposedly required to alter carbon isotope values compared to those of oxygen isotope values (Banner and Hanson 1990).

Carbon isotope excursion and possible causes

The C and O isotopic data generated for subsurface samples of Late Paleocene to Early Eocene (Table. 1) has been plotted in Fig. 3. The P–E transition (~55 ma) is represented by prominent 2–3‰ negative carbon isotope (CIE). The most consistent geochemical signature recorded about coeval with the PETM is the negative CIE. The asymmetric shape has been interpreted as a geologically rapid input of C-depleted carbon into the system, followed by a gradual sequestration of the excess carbon. A short event (~4‰) that appears similar to the PETM in nature has been observed preceding the PETM in the study area (Fig. 3). Cramer et al. 2003 have also reported several short, negative carbon isotope shifts at deep-sea sites, which resemble the much larger-amplitude carbon isotope excursion at the PETM. Bains et al. 1999 also reported negative steps in bulk carbonate C records locally at the onset of PETM. In general, post-CIE δ^{13}C values appear always lower than pre-CIE values, which may be related to the background late Paleocene–early Eocene decrease in exogenic C (Zachos et al. 2001).

Major perturbations of the global carbon cycle during this transition suggest carbon release by natural processes

Table 1 Carbon and oxygen isotope composition of carbonates representing Paleocene–Eocene transition in Jaisalmer Basin, India

	Sample	Formation	$\delta^{13}C_{PDB}$‰	$\delta^{18}O_{PDB}$‰	$\delta^{18}O_{SMOW}$‰
Eocene	835–830	Khuiala	−0.20	−4.89	25.82
	860–855		−0.17	−4.79	25.93
Paleocene	885–880	Sanu	−1.99	−5.61	25.08
	910–905		−2.21	−6.15	24.52
	935–930		−3.59	−6.93	23.71
	980–975		−3.53	−7.70	22.92
	985–980		−0.83	−5.39	25.31
	990–985		−0.78	−5.41	25.28
	1,005–1,000		−0.36	−5.68	25.01
	1,040–1,005		−0.03	−6.23	24.44
	1,045–1,040		−4.13	−7.67	22.96
	1,050–1,045		−0.28	−4.90	25.81
	1,055–1,050		0.97	−5.28	25.41
Cretaceous	1,060–1,055	Parh	−1.23	−5.62	25.06
	1,065–1,060		−0.99	−5.52	25.17
	1,070–1,065		−0.26	−4.88	25.83
	1,075–1,070		0.07	−4.54	26.18

on timescales (Dickens 1999). The leading hypothesis to explain the extreme greenhouse conditions prevalent during this period is the dissociation of 1,400–2,800 giga tonnes of methane from ocean clathrates (Dickens et al. 1997), resulting in a large negative carbon isotope excursion and severe carbonate dissolution in marine sediments. The global warming began during the mid-Paleocene and continued across the P–E transition, peaking during the early Eocene climatic optimum (Zachos et al. 1994, 2001). A short negative carbon isotope excursion event preceding the PETM in the study area (Fig. 3) suggests that warming trend is coincident with a long-term decrease of carbon isotopic values that began ~57 Ma. Several short, negative carbon isotope shifts at deep-sea sites resemble the much larger-amplitude carbon isotope excursion at the PETM (Cramer et al. 2003; Sluijs et al. 2007).

It may be observed from Fig. 2 that significant PETM-related transgression began shortly before the globally recorded negative carbon isotope excursion. Transgressions within such little time are unlikely to have been caused by tectonic forcing. Coupled ocean–climate model simulations have recently indicated that even with 4–8× pre-industrial CO concentrations in the atmosphere, small ice sheets were possibly present at high altitudes on the Antarctica during the late Paleocene. Such models predict that melting of such ice sheets may have contributed to 5–10 m of sea level rise. In addition, thermal expansion of sea water as a result of the ~5°C warming of the ocean likely contributed in a similar magnitude to the sea level rise across the PETM. The idea that warming occurred simultaneously with the CIE at the PETM derives from the numerous deep-sea carbonate $\delta^{13}C$ and $\delta^{18}O$ records, which show negative excursions at the same stratigraphic level. Neritic sections

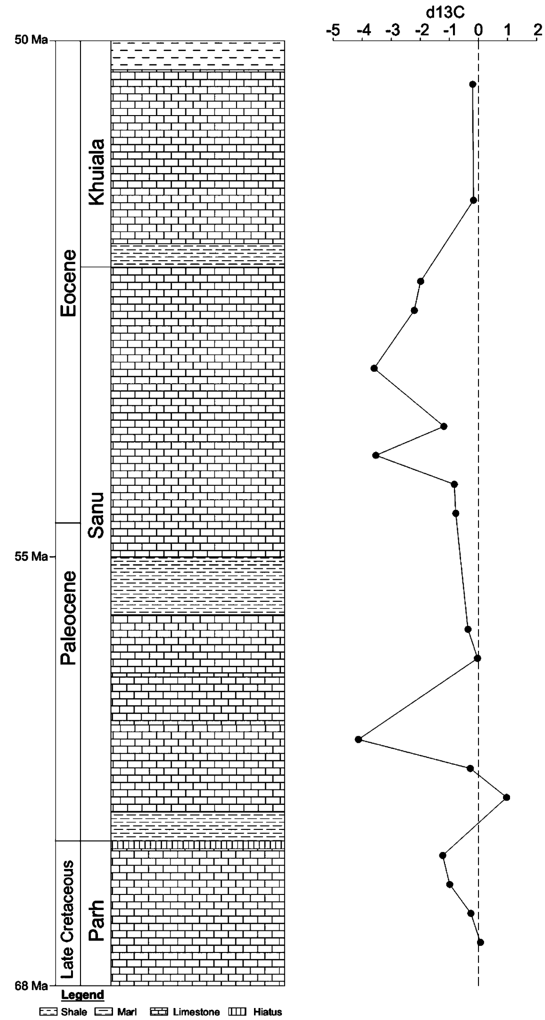

Fig. 3 Secular variations in $\delta^{13}C$ of carbonates from Paleocene–Eocene transition, Jaisalmer Basin

potentially yield much higher sedimentation rates, particularly across the PETM, because the transgression resulted in larger accommodation space on the shelves.

A possible mechanism for the warming of climate for this event may include explosive volcanism (Bralower et al. 1997; Schmitz et al. 2004), comet impact (Kent et al. 2003) and plate tectonics processes. Substantial basaltic volcanism "Deccan traps" had been active in western India around the preceding million years to the P–E transition. The volcanism that initiated in Late Cretaceous continued until the lower part of the upper Paleocene. The Tertiary sediments in adjoining sedimentary basins (Kutch and Cambay Basin) deposited on these Deccan traps magmatism. Intrusions of hot magma into carbon-rich sediments may have triggered the degassing of isotopically light methane in sufficient volumes to cause global warming and the observed isotope anomaly. The climate of the late Paleocene through early Eocene followed a clear long-term warming trend, as evidenced by benthic foraminifer $\delta^{18}O$

(Zachos et al. 2001). This warming is potentially related to increasing CO_2 levels through high volcanic activity in the North Atlantic Igneous Province (Maclennan and Jones 2006; Schmitz et al. 2004; Thomas and Bralower 2005) and along Indian Ocean spreading zones (Cogne and Humler 2006).

A briefly popular theory held that a ^{12}C-rich comet struck the earth and initiated the warming event. This would require a catastrophic impact and which in turn should have left its mark on the globe. Spherules, glassy balls and highly magnetic fine dust have been reported from the bone bed of the Fatehgarh Formation (Late Cretaceous) of the adjoining Barmer Basin (Mathur et al. 2005a, b) In addition, plate tectonics was very active in this part of subcontinent during the late Paleocene–early Eocene interval (−57.4 to 53.4 Ma). India–Asia collision has been recognized as a significant tectonic process during this transition in the Indian subcontinent. It appears that the pre-CIE global changes somehow triggered the injection of ^{12}C- enriched carbon. The timescale for thermal destabilization of methane hydrates is in the order of thousands of years, Hence, sea surface conditions characteristic of PETM, including extreme warming, initiated significantly prior to the injection of ^{12}C-enriched carbon. This implies that this injection likely occurred as a result of global change, rather than the other way around. Full understanding of the Paleocene–Eocene transition in Jaisalmer Basin will require investigation at even finer scales of temporal resolution, and results reported here suggest that finer-scale studies should be possible.

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