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Latest Danian carbon isotope anomaly and associated environmental change in the southern Tethys (Nile Basin, Egypt)

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Abstract: During the Palaeocene and Eocene Epochs (65.5–33.9 Ma) the Earth experienced the warmest conditions of the Cenozoic. This Palaeogene greenhouse episode is characterized by several short-lived negative carbon isotope ($\delta^{13}\text{C}$) excursions, which are usually interpreted as transient warming events ('hyperthermals') as indicated by rising temperatures of surface and bottom waters, and accompanied carbonate dissolution in deep-sea settings. Among these events, the Palaeocene–Eocene Thermal Maximum (PETM, c. 55.5 Ma) is globally the best documented and most prominent. Further negative $\delta^{13}\text{C}$ anomalies have recently been identified in the Eocene, but the early to middle Palaeocene has mostly been neglected. Here, benthic foraminiferal $\delta^{13}\text{C}$ records are presented from four upper Danian–lower Selandian sections in the Nile Basin (eastern Egypt). All records show a negative $\delta^{13}\text{C}$ shift with an amplitude of up to 2‰ at the base of planktonic foraminiferal subzone P3b (c. 61 Ma). The supra-regional nature of this event is emphasized by correlation with a well-dated, similar $\delta^{13}\text{C}$ record from Zumaia (Spain). Lithological changes, the $\delta^{13}\text{C}$ signature and biotic responses strongly resemble those of the PETM in Egypt, which leads to the hypothesis that this 'Latest Danian Event' (LDE) may represent another early Palaeocene hyperthermal.

Several negative carbon isotope ($\delta^{13}\text{C}$) excursions ($>0.5\text{‰}$) have been identified during the Palaeogene greenhouse episode (e.g. Kennett & Stott 1991; Cramer *et al.* 2003; Lourens *et al.* 2005; Nicolo *et al.* 2007; Quillévéré *et al.* 2008). The most prominent, and probably most severe one is the Palaeocene–Eocene Thermal Maximum (PETM) at c. 55.5 Ma, which shows a rapid $\delta^{13}\text{C}$ decrease of $>2.5\text{‰}$ and lasted for about 170–220 ka (Kennett & Stott 1991; Bralower *et al.* 1995; Röhl *et al.* 2000, 2007; Zachos *et al.* 2001, 2006). This negative $\delta^{13}\text{C}$ shift coincides with decreasing $\delta^{18}\text{O}$ values of foraminiferal shells, which reflect short-lived hyperthermal conditions indicating a warming of bottom and surface waters of up to 10 °C (Kennett & Stott 1991; Zachos *et al.* 2003, 2006; Tripathi & Elderfield 2005).

To explain the observed negative carbon isotope excursions, it has been postulated (e.g. Dickens *et al.* 1997) that isotopically light carbon was added to the ocean and atmosphere system, perturbing the carbon cycle as reflected in negative $\delta^{13}\text{C}$ values of marine and terrestrial carbonates, and organic matter (e.g. Kennett & Stott 1991; Bowen *et al.* 2004; Pagani *et al.* 2006a; Yans *et al.* 2006). These carbon isotope excursions are usually ascribed to the release of ^{12}C -enriched methane from the sea floor, which oxidized rapidly to produce CO_2 (Dickens *et al.* 1997). The input of huge amounts of CO_2 into the ocean–atmosphere system increased global temperature, accelerated the hydrological and weathering cycle, and induced a substantial shoaling of the ocean lysocline and the calcite compensation depth (e.g. Dickens *et al.* 1997; Ravizza *et al.* 2001; Schmitz *et al.* 2001; Bowen *et al.* 2004; Zachos *et al.* 2005). However, reliable quantification of the amount of released CO_2 and its source is still under discussion (Higgins & Schrag 2006; Pagani *et al.* 2006b).

Further, less pronounced negative $\delta^{13}\text{C}$ inflections have been identified in the Eocene (Bralower *et al.* 2002; Cramer *et al.* 2003; Lourens *et al.* 2005; Nicolo *et al.* 2007). Also, for the Palaeocene several less well-documented candidates for transient warming events have been proposed (Thomas & Zachos 2000), which include: (1) the early Danian Dan-C2 event (c. 65.2 Ma; Quillévéré *et al.* 2008); (2) the Danian–Selandian (D–S) transition (c. 61 Ma; Speijer 2003); (3) the mid-Palaeocene biotic event (c. 58.2 Ma; Bralower *et al.* 2002; Petrizzo 2005; Bernaola *et al.* 2007). To date, evidence for ocean warming in addition to a negative $\delta^{13}\text{C}$ excursion has been reported only for the Dan-C2 event.

This study is focused on the Danian–Selandian transition in the Eastern Desert of Egypt. This area is part of the southern Tethyan margin and represents an important area for studying processes on the Palaeocene continental margin, as it provides numerous well-exposed sections with well-preserved microfossils from a variety of palaeoenvironmental settings (e.g. Speijer & Schmitz 1998; Guasti *et al.* 2006). In particular, Palaeocene sequences in the Nile Basin, Egypt, provide excellent opportunities to identify the signature of global events on biotic and environmental evolution in the subtropical–tropical neritic realm of the southern Tethys (e.g. Speijer & Schmitz 1998; Speijer 2003). Based on the many micropalaeontological and lithological similarities between the D–S transition and the PETM in Nile Basin sequences (Speijer 2003), and through comparison with preliminary benthic foraminiferal data from the Ocean Drilling Program (ODP; Thomas & Zachos 2000), it has been suggested that the D–S transition may record one of the earliest Cenozoic hyperthermals (Speijer 2003).

In addition, the Spanish Zumaia section, which hosts the Global Stratotype and Stratigraphic Point (GSSP) for the Selandian (Schmitz *et al.* 2008), displays two negative $\delta^{13}\text{C}$ excursions: the first, smaller one with an amplitude of *c.* 0.5‰ in the lowermost part of magnetochron C26r, and a second one with an amplitude of *c.* 1‰ just above the lowest occurrence (LO) of *Fasciculithus tympaniformis*; that is, the base of the calcareous nannofossil biozone NP5 (Arenillas *et al.* 2008). However, $\delta^{13}\text{C}$ records for the D–S transition are based either on low-resolution records (Egypt and Israel: Charisi & Schmitz 1998; Wombat Plateau: Quillévéré *et al.* 2002; Quillévéré & Norris, 2003) or on bulk-rock analyses (Zumaia section: Schmitz *et al.* 1997, 1998; Arenillas *et al.* 2008). Here, the first $\delta^{13}\text{C}$ records are presented from the D–S transition based on benthic foraminifers (*Pyramidulina*) from four sections in eastern Egypt (Fig. 1).

Biostratigraphy and lithology

The studied interval covers the planktonic foraminiferal biozones P2–P4 following Berggren & Pearson (2005), and the calcareous nannofossil zones NP4 and NP5 and NTP6–NTP9 after Martini (1971) and Varol (1989), respectively. A detailed discussion of the biostratigraphy has been given by Steurbaut & Sztraikos (2008) and Sprong *et al.* (2009). The Danian and Selandian stages are represented by the upper part of the Dakhla Formation, which is made up of monotonous brownish to grey marls and shales deposited on the generally NW-deepening shelf (Said 1990; Speijer 2003). A prominent centimetre- to decimetre-thick event deposit is intercalated within the marly Dakhla Formation close to the P3a–P3b planktonic foraminiferal subzonal boundary (Sprong *et al.* 2009). The lower contact of the event deposit with the enclosing Dakhla beds is sharp and possibly unconformable (Speijer 2003; Sprong *et al.* 2009).

Two sections have been studied at Gebel Qreiya, named Qreiya 1 and Qreiya 3. Qreiya 1 is located on the western rim of Gebel Qreiya (26°30.188'N, 32°52.305'E), and Qreiya 3 is located near the southern tip of Gebel Qreiya (26°27.702'N, 33°1.905'E). At both sections lower Danian to Ypresian sediments attain a thickness of *c.* 100 m (Said 1990). The palaeobathymetry has been estimated at 150–250 m (Speijer 2003; Sprong *et al.* 2009). The upper Danian event deposits consist of two distinct beds (beds I and II). At Qreiya 1, laminated dark

brown–pinkish marl beds interrupt the sediments of the Dakhla Formation from 4.10 to 4.22 m. The lowermost 2 cm are rich in coprolites and planktonic foraminifers (= upper part bed I), followed up-section by a dark greyish shale of 6 cm thickness. The uppermost 4 cm are again laminated brown marls with clay pebbles, fish remains and abundant planktonic foraminifers. Two metres higher up-section (6.6 m) a channel bed of 5 cm thickness occurs, and from 10.2 m upwards the carbonate content of the marls increases gradually. At Qreiya 3 the base of the upper Danian event beds is positioned at 8.20 m, measured from the base of the wadi floor. The uppermost 15 cm of the grey marl below the event beds are dark grey and bioturbated at the top. Bed I (8.2–8.3 m) consists of a dark pinkish brown, organic-rich laminated marl (up to 3.5 wt% total organic carbon (TOC), L. Schwark, pers. comm.) containing fish remains, coprolites and planktonic foraminifers. The upper 7 cm of bed I are again made of the pinkish brown laminated marls. Parallel to the lamination dark grey clay lenses are present, which represent downward bioturbation from bed II (8.3–8.5 m), which is a grey marly shale and contains hematitic or limonitic bivalve and gastropod moulds.

The Gebel Aweina section (25°14.650'N, 32°45.994'E) is located east of the Nile Valley, about 50 km south of Luxor. The palaeobathymetry of this section has been estimated at 150–250 m (Speijer & Schmitz 1998; Speijer 2003; Sprong *et al.* 2009), probably slightly shallower than the Qreiya sections. The lower Palaeogene succession is more than 150 m thick. The upper Danian event beds are situated *c.* 25 m above the Cretaceous–Palaeogene (K–P) boundary (Speijer & Schmitz 1998) within the marls of the Dakhla Formation and attain a thickness of only 5 cm. The lowermost 3 cm consist of laminated dark brown–pinkish marls filled with coprolites and high numbers of planktonic foraminifers (upper part bed I), whereas the remaining 2 cm consist of a dark grey shale (bed II) with remains of bivalves and gastropods.

The Gebel Duwi section (26°06.093'N, 34°06.707'E) is located about 15 km inland from the Red Sea coast, west of Quseir, and represents the shallowest section studied (70–150 m; Speijer 2003). The studied part spans 12 m of the Dakhla Formation and starts 32 m above the K–P boundary. At 37.6 m the grey marls are interrupted by the event deposit, which is represented by a single light grey sandy marl bed of 10 cm thickness that shows

Fig. 1. Map showing the locations of the sections studied and others mentioned in the text. Egypt inset displays the detailed location of the studied sections, which are located along a palaeobathymetric gradient; projected palaeoisobaths are indicated.

no particular enrichment in organic matter, but fines upwards, with bioturbation and lamination in the lowermost 5 cm. In addition, it also contains fish remains and coprolites.

Methods

Samples were dried for at least 24 h at 60 °C and then soaked in a 50 g l⁻¹ Na₂CO₃ solution for 24 h. $\delta^{13}\text{C}$ analyses were performed on well-preserved specimens of the nodosariid benthic foraminifers *Pyramidulina latejugata* and *P. macneili*. The specimens were picked from the >250 μm fraction, the shells were cracked between two glass slides and the resulting fragments were separated manually from the infilling. Shell material was cleaned with a picking needle and distilled water. The shell carbonate was reacted with 100% phosphoric acid at 75 °C using a Kiel III online carbonate preparation line connected to a ThermoFinnigan 252 mass spectrometer. Reproducibility was checked by replicate analysis of laboratory standards and was better than $\pm 0.05\text{‰}$ and 0.06‰ (1σ) for $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$, respectively. All values are reported in per mil relative to VPDB by assigning a $\delta^{13}\text{C}$ value of +1.95‰ and a $\delta^{18}\text{O}$ value of -2.20‰ to NBS19. Results are shown in Table 1.

Results and discussion

Preservation and life habitat

Preservation of the shell material was evaluated using scanning electron microscopy (Fig. 2). Secondary cements attached to the shells have not been observed, and the ultrastructure shows original shell structure and pore canals (Fig. 2); however, minor signs of incipient recrystallization have been recognized on some specimens. Therefore, the focus will be mainly on the carbon isotope record, as oxygen isotope values are more prone to diagenetic alteration during burial than carbon isotope values (Marshall 1992).

Pyramidulina species have successfully been used for stable isotope studies of Palaeogene (Charisi & Schmitz 1998) and Cretaceous sediments (Almogi-Labin *et al.* 1993; Fassell & Bralower 1999). Stable isotope data (Shackleton *et al.* 1984; Fassell & Bralower 1999) and morphological studies (Koutsoukos & Hart 1990) have suggested that *Pyramidulina* had either an epifaunal (Shackleton *et al.* 1984; Koutsoukos & Hart 1990) or a shallow infaunal lifestyle (Koutsoukos & Hart 1990; Fassell & Bralower 1999). Data of Fassell & Bralower (1999) showed a fairly good agreement between $\delta^{13}\text{C}$ trends in epifaunal gavelinellids and *Pyramidulina*, indicating that this group could be well suited to trace relative changes in bottom water $\delta^{13}\text{C}$.

Late Danian $\delta^{13}\text{C}$ and environmental changes

All four sections studied show a negative shift of at least 1‰ in $\delta^{13}\text{C}$ (Fig. 3), with the most negative values within or above the upper Danian event beds, slightly above the P3a–P3b subzonal boundary as delineated by the LO of *Igorina albeari* (Sprong *et al.* 2009). Calcareous nannofossil studies (Gebel Qreiya and Gebel Aweina) correlate this to a level within zones NP4 and lowermost NTp7B (Sprong *et al.* 2009). At Gebel Duwi, the shallowest section, the excursion seems to be expanded and shows a gradual 2‰ shift, suggesting that the magnitude of this excursion might have been greater than 1‰. At Gebel Duwi the values start to decline about 1 m below the event beds, and the most negative values occur slightly above these beds. In both Qreiya sections a second negative excursion of c. 1.3‰ was

Fig. 2. Scanning electron micrographs of a specimen of the benthic foraminifer *Pyramidulina*: (a) whole test view; (b, c) detail showing well-preserved original shell ultrastructure with vertical pore canals.

found close to an unconformity (Fig. 3) at the top of subzone P3b and slightly below the NP4–NP5 calcareous nannofossil boundary.

There are two feasible scenarios to explain the prominent negative shift in $\delta^{13}\text{C}$ above the P3a–P3b subzonal boundary. One explanation on a local scale would be that isotopically light CO₂, which may be released during degradation of organic matter in the event deposit, has become incorporated into shells of benthic foraminifers, causing a trend in $\delta^{13}\text{C}$ towards negative values. However, this explanation is discarded for three reasons: (1) the $\delta^{13}\text{C}$ excursion is not related to the organic matter rich event deposit in all sections (e.g. at Gebel Duwi; Fig. 3); (2) at Gebel Duwi the $\delta^{13}\text{C}$ values show a decrease below the event bed and minimum values occur above it, and are therefore not caused by the degradation of organic matter; (3) there is ample stratigraphical evidence (Steurbaut & Sztrákos 2008; Sprong *et al.* 2009) that this excursion has a supra-regional nature, as it is also known from northern Spain (Zumaia, 900–1100 m water depth, North Atlantic Ocean; Arenillas *et al.* 2008; Fig. 3).

The second explanation is similar to that suggested for the PETM (e.g. Dickens *et al.* 1997): huge amounts of isotopically light carbon may have been added to the ocean–atmosphere system, possibly as a result of the release and subsequent dissociation of methane, or alternatively through oxidation of terrestrial or marine organic carbon, or enhanced volcanic outgassing (e.g. Pagani *et al.* 2006b). In this case the latest Danian $\delta^{13}\text{C}$ excursion should also occur in other oceanic basins.

In Egypt several palaeoenvironmental characteristics of this Latest Danian Event (LDE) are similar to those of the PETM (Speijer 2003; see Table 2 for a summary). The most prominent similarities are: (1) low levels of bottom-water oxygenation have caused enhanced preservation of organic matter during both events; (2) both event deposits show a very high planktonic/benthic foraminiferal ratio, and the benthic foraminiferal assemblage is dominated by opportunistic taxa indicating shallow water (*Neoponides duwi* in the latest Danian, and *Anomalinoidea*

Table 1. Comparison of sedimentary and biotic characteristics between the Latest Danian Event, the PETM and the background sedimentation (modified from Speijer 2003)

Section	Depth (m)	$\delta^{13}\text{C}$ (‰ PDB)	$\delta^{18}\text{O}$ (‰ PDB)
Gebel Qreiya 1	1.00	-0.39	-2.35
Gebel Qreiya 1	1.50	-0.46	-2.06
Gebel Qreiya 1	3.50	-0.32	-2.03
Gebel Qreiya 1	3.70	-0.70	-2.01
Gebel Qreiya 1	3.95	-0.61	-2.42
Gebel Qreiya 1	4.10	-0.98	-2.84
Gebel Qreiya 1	4.12	-1.22	-2.88
Gebel Qreiya 1	4.17	-0.95	-2.45
Gebel Qreiya 1	4.18	-0.45	-2.48
Gebel Qreiya 1	4.19	-0.97	-2.46
Gebel Qreiya 1	4.22	-0.72	-2.46
Gebel Qreiya 1	4.60	0.00	-2.08
Gebel Qreiya 1	4.90	-0.28	-3.55
Gebel Qreiya 1	5.50	-0.12	-1.90
Gebel Qreiya 1	5.50	-0.26	-2.95
Gebel Qreiya 1	6.00	-1.40	-1.93
Gebel Qreiya 1	6.70	-0.82	-3.63
Gebel Qreiya 1	8.20	-0.14	-2.03
Gebel Qreiya 1	8.70	-0.59	-2.74
Gebel Qreiya 1	9.20	-0.34	-2.71
Gebel Qreiya 1	9.70	-0.21	-2.16
Gebel Qreiya 3	0.00	-0.18	-1.72
Gebel Qreiya 3	0.00	-0.36	-1.81
Gebel Qreiya 3	0.24	-0.14	-2.47
Gebel Qreiya 3	0.72	-0.18	-1.99
Gebel Qreiya 3	0.97	-0.22	-1.51
Gebel Qreiya 3	1.45	-0.28	-1.55
Gebel Qreiya 3	1.69	-0.26	-1.89
Gebel Qreiya 3	2.17	-0.17	-1.43
Gebel Qreiya 3	2.41	0.44	-1.45
Gebel Qreiya 3	4.34	-0.11	-2.38
Gebel Qreiya 3	4.34	-0.16	-2.32
Gebel Qreiya 3	4.34	-0.48	-2.28
Gebel Qreiya 3	5.31	-0.29	-2.09
Gebel Qreiya 3	5.55	-0.29	-1.94
Gebel Qreiya 3	6.28	-0.65	-1.37
Gebel Qreiya 3	7.00	-0.01	-1.47
Gebel Qreiya 3	7.50	0.00	-1.56
Gebel Qreiya 3	8.00	-0.94	-2.24
Gebel Qreiya 3	8.32	-0.97	-1.90
Gebel Qreiya 3	8.35	-1.28	-1.74
Gebel Qreiya 3	8.42	-0.43	-1.59
Gebel Qreiya 3	8.50	-0.90	-2.24
Gebel Qreiya 3	9.21	-1.06	-1.54
Gebel Qreiya 3	9.45	-0.48	-1.56
Gebel Qreiya 3	9.92	-0.91	-1.38
Gebel Qreiya 3	9.92	-0.77	-1.54
Gebel Qreiya 3	11.10	0.05	-1.07
Gebel Qreiya 3	11.34	-0.44	-2.05
Gebel Qreiya 3	11.57	-0.36	-2.04
Gebel Qreiya 3	11.81	-0.05	-1.74
Gebel Qreiya 3	12.05	-0.57	-2.09
Gebel Qreiya 3	12.28	-0.54	-1.95
Gebel Qreiya 3	12.75	-0.46	-2.21
Gebel Qreiya 3	13.23	-0.72	-1.84
Gebel Qreiya 3	13.70	-0.85	-2.33
Gebel Qreiya 3	13.94	0.27	-2.11
Gebel Qreiya 3	15.44	0.19	-1.79
Gebel Aweina 95	19.00	-0.46	-2.33
Gebel Aweina 95	22.00	-0.06	-2.08
Gebel Aweina 95	23.00	0.05	-1.93
Gebel Aweina 95	25.45	-0.56	-2.33
Gebel Aweina 95	25.68	-1.18	-2.09
Gebel Aweina 95	25.79	-1.03	-2.17
Gebel Aweina 95	25.88	-1.76	-2.44
Gebel Aweina 95	25.94	-1.47	-2.19
Gebel Aweina 95	25.98	-1.87	-2.51
Gebel Aweina 95	26.00	-0.62	-2.45

(continued)

Table 1. (continued)

Section	Depth (m)	$\delta^{13}\text{C}$ (‰ PDB)	$\delta^{18}\text{O}$ (‰ PDB)
Gebel Aweina 95	26.03	-1.05	-2.86
Gebel Aweina 95	26.07	-1.15	-2.45
Gebel Aweina 95	26.11	-1.59	-2.50
Gebel Aweina 95	26.22	-1.34	-2.53
Gebel Aweina 95	26.38	-1.03	-2.11
Gebel Aweina 95	26.48	-0.63	-2.28
Gebel Aweina 95	26.58	-0.71	-1.95
Gebel Aweina 95	26.68	-0.93	-2.02
Gebel Aweina 95	26.95	-0.82	-2.55
Gebel Aweina 95	27.25	-0.54	-2.24
Gebel Aweina 95	27.75	-0.95	-2.59
Gebel Aweina 95	28.25	-0.85	-2.56
Gebel Aweina 95	28.50	-0.33	-2.37
Gebel Aweina 95	28.75	-0.35	-1.91
Gebel Aweina 95	29.00	0.12	-2.06
Gebel Aweina 95	31.00	-0.61	-2.31
Gebel Aweina 95	34.00	0.04	-1.96
Gebel Aweina 95	35.00	-0.37	-2.52
Gebel Aweina 06	25.75	-1.16	-2.55
Gebel Aweina 06	25.85	-0.87	-2.57
Gebel Aweina 06	25.90	-1.21	-2.66
Gebel Aweina 06	25.95	-1.28	-2.80
Gebel Aweina 06	25.99	-0.54	-2.64
Gebel Aweina 06	26.01	-0.61	-2.06
Gebel Aweina 06	26.02	-0.93	-2.52
Gebel Aweina 06	26.05	-0.56	-2.23
Gebel Aweina 06	26.10	-0.84	-2.23
Gebel Aweina 06	26.15	-0.78	-2.64
Gebel Aweina 06	26.25	-0.50	-2.48
Gebel Duwi 95	30.07	-0.08	-1.68
Gebel Duwi 95	31.01	-0.34	-1.49
Gebel Duwi 95	31.78	-0.22	-1.94
Gebel Duwi 95	32.57	0.00	-1.61
Gebel Duwi 95	33.34	0.16	-1.75
Gebel Duwi 95	34.12	-0.28	-1.77
Gebel Duwi 95	34.91	-0.15	-1.58
Gebel Duwi 95	35.70	-0.02	-1.41
Gebel Duwi 95	36.64	-0.14	-1.62
Gebel Duwi 95	37.58	-1.32	-2.46
Gebel Duwi 95	38.05	-0.52	-1.59
Gebel Duwi 95	38.05	-0.47	-1.41
Gebel Duwi 95	38.52	-0.67	-1.61
Gebel Duwi 95	40.40	-0.09	-2.39
Gebel Duwi 95	41.34	-0.99	-1.27
Gebel Duwi 95	42.28	-0.08	-1.47
Gebel Duwi 95	43.22	0.06	-1.61
Gebel Duwi 06	34.11	0.15	-1.73
Gebel Duwi 06	34.51	0.11	-1.49
Gebel Duwi 06	34.98	-0.44	-2.20
Gebel Duwi 06	35.38	-0.59	-2.10
Gebel Duwi 06	35.78	-0.35	-1.78
Gebel Duwi 06	36.18	-0.19	-1.69
Gebel Duwi 06	36.58	-0.32	-2.18
Gebel Duwi 06	36.98	-0.33	-2.28
Gebel Duwi 06	37.18	0.15	-1.98
Gebel Duwi 06	37.38	-0.49	-1.79
Gebel Duwi 06	37.43	-0.23	-2.07
Gebel Duwi 06	37.48	-0.32	-1.87
Gebel Duwi 06	37.57	-0.76	-1.73
Gebel Duwi 06	37.58	-0.83	-2.14
Gebel Duwi 06	37.66	0.07	-1.70
Gebel Duwi 06	37.71	-0.98	-1.90
Gebel Duwi 06	37.76	-0.77	-1.73
Gebel Duwi 06	37.81	-0.49	-1.40
Gebel Duwi 06	37.83	-1.46	-1.93
Gebel Duwi 06	37.98	-0.51	-1.50
Gebel Duwi 06	38.18	-2.05	-1.95
Gebel Duwi 06	38.38	-1.50	-2.14
Gebel Duwi 06	38.78	-0.84	-2.03
Gebel Duwi 06	39.18	-1.56	-2.52

Fig. 3. Lithology, stratigraphy and benthic foraminiferal $\delta^{13}\text{C}$ records from four sections (Gebel Qreiya 1, Gebel Qreiya 3, Gebel Aweina, Gebel Duwi) in the Nile Basin (Egypt), and their correlation with the Selandian GSSP succession at Zumaia (Spain, Arenillas *et al.* 2008) and ODP Hole 761B (offshore NW Australia, Quillévéré *et al.* 2002; Quillévéré & Norris 2003). The P3a–P3b planktonic foraminiferal subzonal boundary (*) at Zumaia has been changed with respect to Arenillas *et al.* (2008). Taxonomic comparisons reveal that *Igorina cf. albeari* (also named *Morozovella crosswicksensis*) of Arenillas *et al.* (2008) is identical to our concept of *I. albeari* and the P3a–b subzonal boundary has consequently been moved downwards (see also discussion by Sprong *et al.* 2009). Grey shaded area represents the correlation between Zumaia and the Egyptian section for the Latest Danian Event; the dashed grey band shows a tentative correlation for the upper $\delta^{13}\text{C}$ excursion at Zumaia.

Table 2. Comparison of sedimentary and biotic characteristics between the Late Danian Event, the PETM and the background sedimentation (modified from Speijer 2003)

Parameter	Late Danian Event	PETM	Palaeocene background
<i>Lithology/geochemistry</i>			
Predominant lithology	Dark brown marl	Dark brown marl	Grey to pale brown marl
Discontinuities	Unconformity	Some distinct	Some paraconformities
Systems tract	Transgressive	Transgressive	Various
Fish remains	Abundant	Abundant	Rare
TOC (%)	0.8–3.5	1.5–2.7	<0.3
$\delta^{13}\text{C}$ trends	Negative excursion $\leq 2\text{‰}$	Negative excursion $\leq 3\text{‰}$	Long-term variation
<i>Biota/foraminifers</i>			
Planktonic foraminifers >125 μm (%)	2–99.8	c. 99	70–95
<i>Morozovella</i>	Rare	Rare	Common to abundant
<i>Praemurica</i> and <i>Acarinina</i>	Dominant	Dominant	Common to abundant
Benthic assemblage	Oligotaxic	Oligotaxic	Diverse
Dominant benthic taxon	<i>Neoeponides duwi</i>	<i>Anomalinoidea aegyptiacus</i>	Depth dependent
Oxygenation at sea floor	Poor	Poor (variable?)	Good to moderate

aegyptiacus during the PETM); (3) the upper Danian event beds represent an unconformity (probably Sel-1 after Hardenbol *et al.* 1998) followed by a sea-level rise, which is similar to the sea-level scenario suggested for the PETM deposits in Egypt (Speijer & Wagner 2002).

The $\delta^{18}\text{O}$ data, which are reported in Table 1, show no unequivocal shift that supports bottom-water warming during the LDE that would make a strong case for a hyperthermal event. In general, the data are very noisy with no distinct pattern. This is not unique for studies on the southern Tethyan margin. Even for the well-known PETM, no $\delta^{18}\text{O}$ record exists from this area that shows a distinctive change towards lighter $\delta^{18}\text{O}$ values of foraminiferal calcite. The absence of such an isotope shift may result from various factors: in contrast to the deep sea, where bottom-water temperatures and continental ice volume are the main driving factors, shelf sediments are also influenced by evaporation–precipitation changes as well as by river discharge, which both may bias the $\delta^{18}\text{O}$ values. In addition, a diagenetic influence can never be ruled out if sediments crop out on the Earth's surface, and as pointed out above, some shells show signs of incipient diagenesis. Nevertheless, the similarities between the PETM and the LDE with respect to their palaeoenvironmental consequences (Table 2) suggest that the mechanisms that caused the two events were very similar or even identical. This conclusion leads to the hypothesis that the LDE represents another Cenozoic hyperthermal event, which is supported by the supra-regional occurrence of the event.

Supra-regional significance of the latest Danian negative $\delta^{13}\text{C}$ excursion

Whole-rock $\delta^{13}\text{C}$ data from Zumaia (Schmitz *et al.* 1997, 1998; Arenillas *et al.* 2008) show two negative excursions of up to 1‰ at a stratigraphical level that can be confidently correlated with Gebel Qreiya 1 and 3 (Fig. 3). The first excursion is situated in the basal metres of the Itzurun Formation, and the second one is located *c.* 10 m lower at the transition from limestone to limestone–marl couplets at the top of the Danian Aitzgorri Limestone Formation. The latter lithological change and carbon isotope excursion correlate biostratigraphically with the upper Danian event beds at the P3a–P3b planktonic foraminiferal boundary, and correspond approximately to the NTp7A–B calcareous nannofossil boundary (Steurbaut & Sztrákös 2008; Sprong *et al.* 2009).

Sample resolution from the Egyptian sections is too low to allow the testing of orbital pacing for the latest Danian $\delta^{13}\text{C}$ event, as observed for other younger hyperthermals (e.g. Cramer *et al.* 2003; Lourens *et al.* 2005). Furthermore, the existence of an unconformity in at least some of the sections studied hampers an estimation of the timing of this event. Here the calculation of the duration of this $\delta^{13}\text{C}$ event is based on the biostratigraphical correlation between Zumaia and Egypt. At Zumaia average sedimentation rates of 1.33 cm ka^{-1} for magnetochrons C27n (1.36 cm ka^{-1}) and C26r (1.29 cm ka^{-1}) have been calculated by Dinares-Turell *et al.* (2007), resulting in a total duration of 191 ka for the $\delta^{13}\text{C}$ anomaly (Fig. 3), which is comparable with the duration of the PETM (Röhl *et al.* 2000, 2007).

Further anomalies at the top of magnetochron C27n and above were identified by Westerhold *et al.* (2008), who found two prominent peaks in Fe intensities from X-ray fluorescence scanning and magnetic susceptibility ('Top Chron 27n Event') at various ODP sites on Walvis Ridge, Shatsky Rise and in the Caribbean Sea, which are similar to the PETM interval. These

two peaks correspond to the two short eccentricity cycles 38 and 39 according to the astronomical Palaeocene time scale of Westerhold *et al.* (2008). Only one of these peaks occurs above Top C27n magnetochron as the LDE at Zumaia (Fig. 3). This suggests a duration of $\leq 100\text{ ka}$ for the LDE, in contrast to the cyclostratigraphic results from Zumaia. The C27n–C26r transition is further characterized by a significant increase in oceanic spreading rates (Westerhold *et al.* 2008) and enhanced volcanic activity at the SE Greenland margin (Sinton & Duncan 1998), which might be related to the palaeoenvironmental changes reported for the latest Danian by Speijer (2003) and herein.

Whereas the first negative $\delta^{13}\text{C}$ excursion is well constrained, the second one is represented by only a few spot measurements. Therefore it is hard to make a case for a global $\delta^{13}\text{C}$ excursion, although a second negative $\delta^{13}\text{C}$ excursion is documented at Zumaia. According to Westerhold *et al.* (2008), this anomaly should correspond to the short eccentricity cycles 48 and/or 49, which show no distinctive change in the Fe intensities.

A negative $\delta^{13}\text{C}$ shift of *c.* 0.6‰ in benthic and deep-dwelling planktonic foraminifers has also been reported by Quillévéré *et al.* (2002) and Quillévéré & Norris (2003) from the Wombat Plateau (Indian Ocean, ODP Hole 761B), which was followed by a slight (2–3 °C) surface-water warming (Fig. 3). The late Danian $\delta^{13}\text{C}$ excursion at the Wombat Plateau comprises the upper part of magnetochron C27n and the lowermost part of magnetochron 26r; however, benthic foraminifers display the most negative numbers after a *c.* 0.5 Ma hiatus (P3a, NP4a) above magnetochron C27n. It is reasonable to consider that this excursion might represent the same event as in the Egyptian sections and the lower one at Zumaia, which is situated a short distance above the C27n–C26r boundary (Arenillas *et al.* 2008). Although Quillévéré & Norris (2003) followed the same taxonomic criteria for *I. albeardi* as employed here (F. Quillévéré, pers. comm.), they assigned the $\delta^{13}\text{C}$ shift to biozone P3a. It is proposed that the large distance between the present sections and the Wombat Plateau, and their different environmental settings, make a slight diachrony in the first appearance of *I. albeardi* likely, so the magnetostratigraphic age assignment is preferred to biostratigraphy for long-distance correlation purposes.

In conclusion, because of the supra-regional nature and the associated palaeoenvironmental changes that are similar to those for the PETM in Egypt (Speijer 2003; Westerhold *et al.* 2008), it is hypothesized that the negative $\delta^{13}\text{C}$ excursion at the P3a–P3b boundary may represent an additional Palaeocene hyperthermal, although reliable temperature data remain lacking for this interval. Various approaches to estimate the duration of the Latest Danian Event suggest that the carbon isotope anomaly lasted between 100 and 191 ka. Until now, age-equivalent deep-sea sections have been sparsely studied and await further investigation to explore the potential global nature of the event.

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