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Main Deccan volcanism phase ends near the K–T boundary: Evidence from the Krishna–Godavari Basin, SE India

G. Keller [a,*], T. Adatte [b], S. Gardin [c], A. Bartolini [c], S. Bajpai [d]

[a] *Geosciences, Princeton University, Princeton NJ 08540, USA*
[b] *Geological Institute, University of Neuchatel, Neuchatel, CH-2007, Switzerland*
[c] *CNRS-UMR 5143, Université Pierre et Marie Curie Paris 6, 4 Place Jussieu, 75252 Paris, France*
[d] *Department of Earth Sciences, Indian Institute of Technology, Roorkee 247 667 Uttarakhand, India*

Abstract

Recent studies indicate that the bulk (80%) of the Deccan trap eruptions occurred over less than 0.8 m.y. in magnetic polarity C29r spanning the Cretaceous–Tertiary (K–T) boundary. Determining where within this major eruptive phase the K–T mass extinction occurred has remained problematic. For this reason, models estimating the biotic and environmental consequences have generally underestimated the rate and quantity of Deccan gas emissions by orders of magnitude leading to conclusions that volcanism could not have been one of the major causes for the K–T mass extinction. In this study we report that the most massive Deccan trap eruption occurred near the K–T mass extinction.

These results are based on sedimentologic, microfacies and biostratigraphic data of 4–9 m thick intertrappean sediments in four quarry outcrops in the Rajahmundry area of the Krishna–Godavari Basin of southeastern India. In this area two Deccan basalt flows, known as the Rajahmundry traps, mark the longest lava flows extending 1500 km across the Indian continent and into the Bay of Bengal. The sediments directly overlying the lower Rajahmundry trap contain early Danian planktic foraminiferal assemblages of zone P1a, which mark the evolution in the aftermath of the K–T mass extinction. The upper Rajahmundry trap was deposited in magnetic polarity C29n, preceding full biotic recovery. These results suggest that volcanism may have played critical roles in both the K–T mass extinction and the delayed biotic recovery.

Keywords: Deccan volcanism; K–T mass extinction; paleoenvironment; Rajahmundry

1. Introduction

Most of the Deccan volcanic province erupted during a period of less than 1 m.y. spanning the Cretaceous–Tertiary (K–T) boundary with an early pulse of volcanism between 67 and 68 Ma (McLean, 1985; Courtillot et al., 1986, 1988; Duncan and Pyle, 1988; Mitchell and Widdowson, 1991; Vandamme and Courtillot, 1992; Widdowson et al., 2000; Hofmann et al., 2000; Sheth et al., 2001). Recent studies show that the bulk of Deccan volcanism (~80%) encompassing up to 3500 m of lava flows, erupted during less than 800 ky in magnetic polarity C29r (64.8–65.6 Ma) (Chenet et al., 2007, in press; Saunders et al., 2007; Jay and Widdowson, 2008) and was likely a major contributor to the K–T mass extinction. However, to date the only link to the K–T boundary is the report of a small iridium anomaly from Deccan intertrappean beds in Anjar, Kutch Province (Bhandari et al., 1995, 1996; Courtillot et al., 2000; Bajpai and Prasad, 2000), which may be of volcanic origin (Hansen et al., 2001; Sant et al., 2003; Shrivastata and Ahmad, 2005). A direct link between Deccan volcanism and the mass extinction has remained elusive due to the lack of intertrappean marine sediments with age diagnostic microfossils. A search for such sediments in the Deccan large igneous province (LIP) has remained futile because intertrappean beds were largely deposited in terrestrial, fluvio-lacustrine and estuarine environments (Khosla and Sahni,

* Corresponding author.
E-mail address: gkeller@princeton.edu (G. Keller).

2003; Cripps et al., 2005). Only in more distant areas of southeastern India (i.e., Krishna–Godavari and Cauvery Basins) can thick units of intertrappean marine sediments be found that permit biostratigraphic age determinations (Raju et al., 1991; Jaiprakash, 1993; Raju et al., 1996).

The most promising area is in Rajahmundry, about 1500 km to the southeast of the main Deccan LIP where a series of lava flows, known as the lower and upper Rajahmundry traps, are exposed in the Krishna–Godavari (KG) Basin and extend about 70 km offshore into the Bay of Bengal (Fig. 1) (Baksi et al., 1994; Raju et al., 1996; Knight et al., 2003). The Rajahmundry traps have been historically considered as part of the original Deccan volcanic province with lava flows traveling along existing river valleys (Venkaya, 1949; Bhimasankaram, 1965). This has been confirmed by magnetostratigraphy and geochemical similarities with the main Deccan volcanic province to the west (Duncan and Pyle, 1988; Lightfoot et al., 1990; Subbarao and Pathak, 1993; Banerjee et al., 1996; Baksi, 2001; Jay and Widdowson, 2008). The Rajahmundry lava flows are thus the furthest traveled in the Deccan succession and perhaps the longest on the planet. They represent part of the Deccan volcanic acme, though not necessarily its peak.

The best age control for the Rajahmundry traps to date is based on magnetic polarity data. The upper trap is in normal polarity C29n and the lower trap in reversed polarity C29r, which spans the K–T boundary (KTB) (Duncan and Pyle, 1988; Subbarao and Pathak, 1993; Knight et al., 2003; Chenet et al., 2007). $^{40}K/^{40}Ar$ and $^{40}Ar/^{39}Ar$ ages yield a calculated mean age of 65.0 ± 0.3 Ma for the main Deccan phase (Hofmann et al., 2000; Chenet et al., 2007), consistent with the K–T boundary age of 65.0 Ma (Cande and Kent, 1991) as used in this study, but not with the proposed astronomical time scale that places the K–T boundary at 65.5 ± 0.3 Ma (Gradstein and Ogg, 2004).

$^{40}K/^{40}Ar$ and $^{40}Ar/^{39}Ar$ ages can thus determine the duration of the main Deccan eruptive phase, but the large error margin prevents identifying the position of the KTB within the Deccan lava pile. Locating the precise stratigraphic position of the KTB within the Deccan LIP is of critical importance in any debate concerning the role of Deccan volcanism in the K–T mass extinction, including the short- and long-term biotic effects and the pre- and post-K–T climatic changes. Similarly important is the timing of the Deccan LIP with respect to the Chicxulub impact, which now appears to have predated the K–T mass extinction by about 300,000 yr (Keller et al., 2003, 2004, 2007).

The Cretaceous–Tertiary transition in the Rajahmundry area of the Krishna–Godavari (KG) Basin is critical to this debate (Fig. 1). The precise stratigraphic position of the KTB here has remained speculative. Ostracod faunas from the intertrappean beds yielded a Danian age (Bhandari, 1995; Khosla and Nagori, 2002). Studies based on planktic foraminifera from the ONGC (Oil and Natural Gas Corporation) Palakollu-A well show a

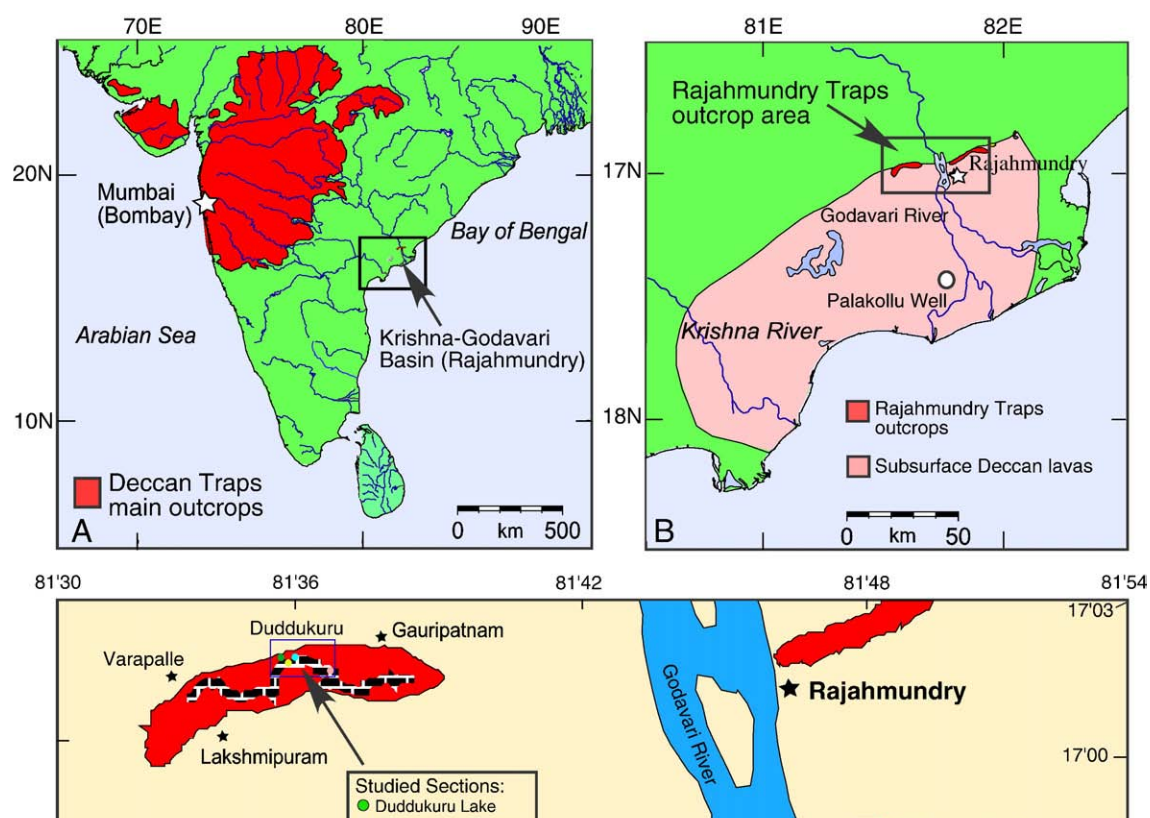

Fig. 1. (A) Present extent of Deccan Traps volcanic province based on outcrop data. (B) Known extent of subsurface Deccan lava with quarry outcrops near the town of Rajahmundry. (C) Surface exposures of the Rajahmundry traps and intertrappean sediments with locations of sections studied. (Modified after Knight et al., 2003).

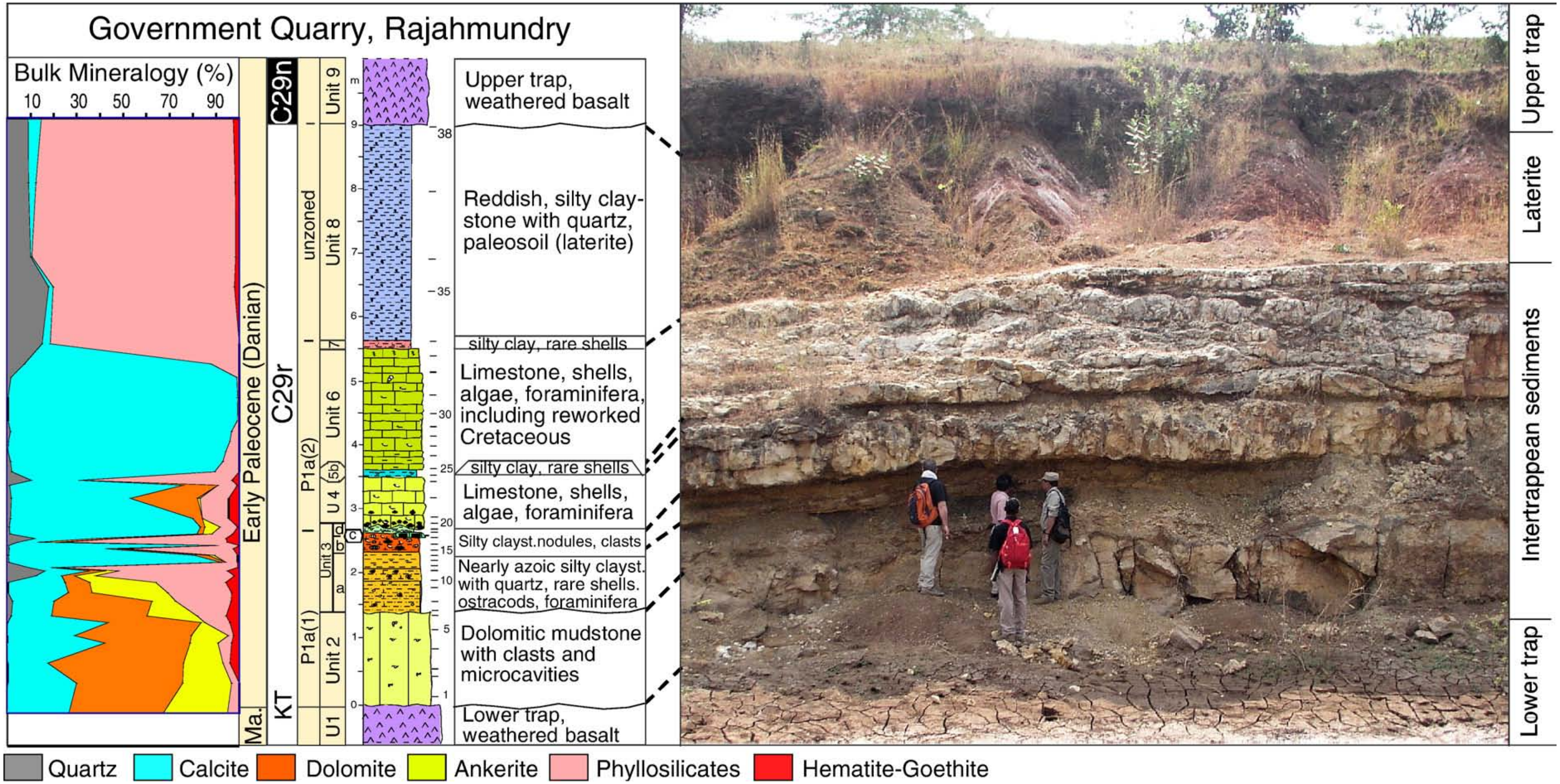

Fig. 2. Government quarry intertrappean sediments span the early Danian Zone P1a. Lithology and bulk rock compositions show deposition occurred in shallow marine to estuarine and subaerial environments. High abundance of phyllosilicates and paleosols marks subaearial environments, high calcite with marine fossils indicates marine deposition. Floodplain or supratidal environments are indicated by high dolomite, ankerite, phyllosilicates and lower calcite.

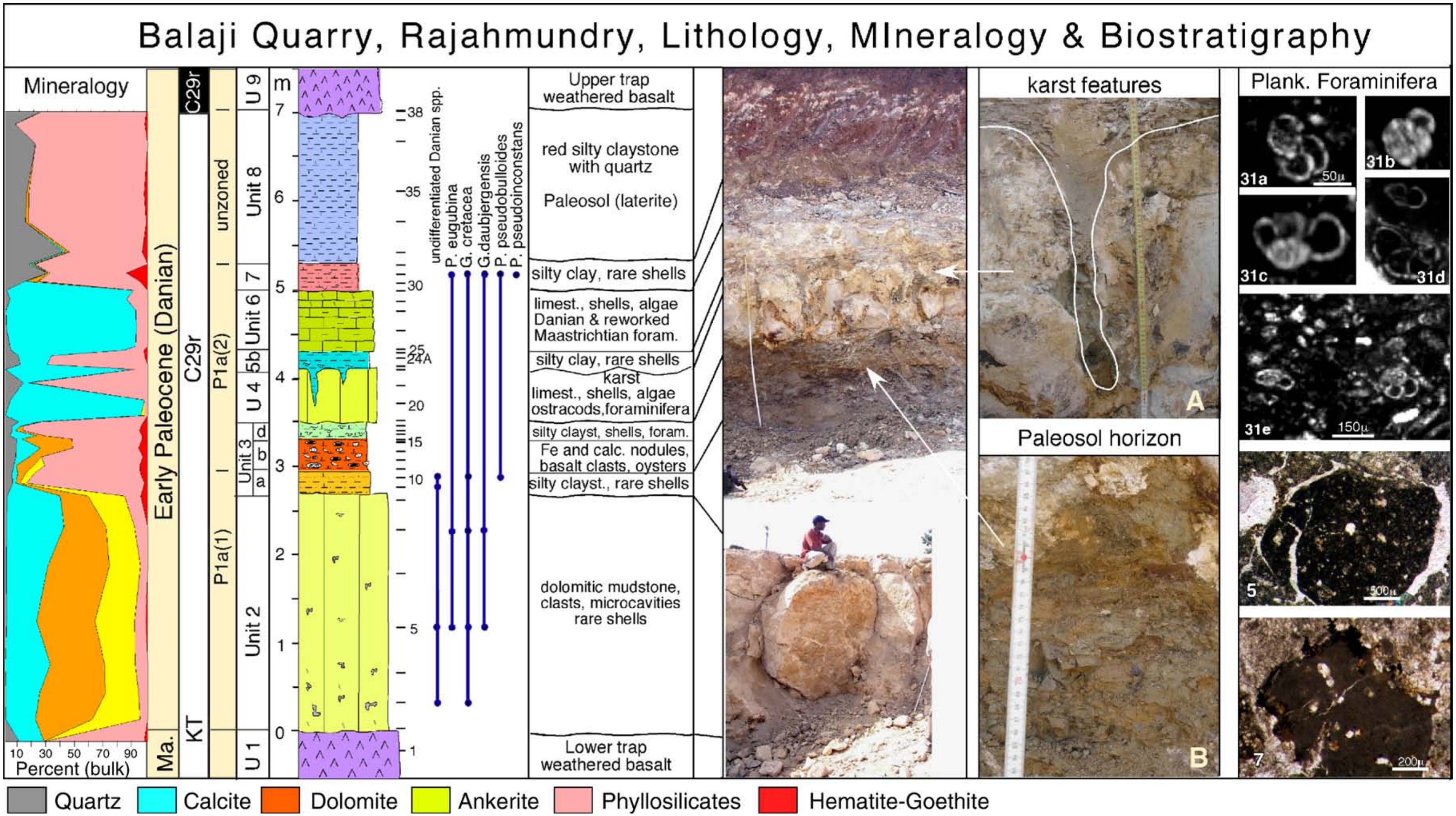

Fig. 3. Lithology, bulk rock mineralogy and biostratigraphy of the Balaji Quarry intertrappean sediments reflect shallow marine to estuarine and subaerial environments similar to Government quarry. (A) karst features, (B) paleosol. Photomicrographs of species (numbers keyed to samples in lithologic column): 5, 7, unit 2 claystone clasts with tiny early Danian species; 31a–e, early Danian species: *Globoconusa daubjergensis* (31a), *Parvularugoglobigerina eugubina* (31b), *Parasubbontina pseudobulloides* (31c,e), *Praemurica compressa* (31d).

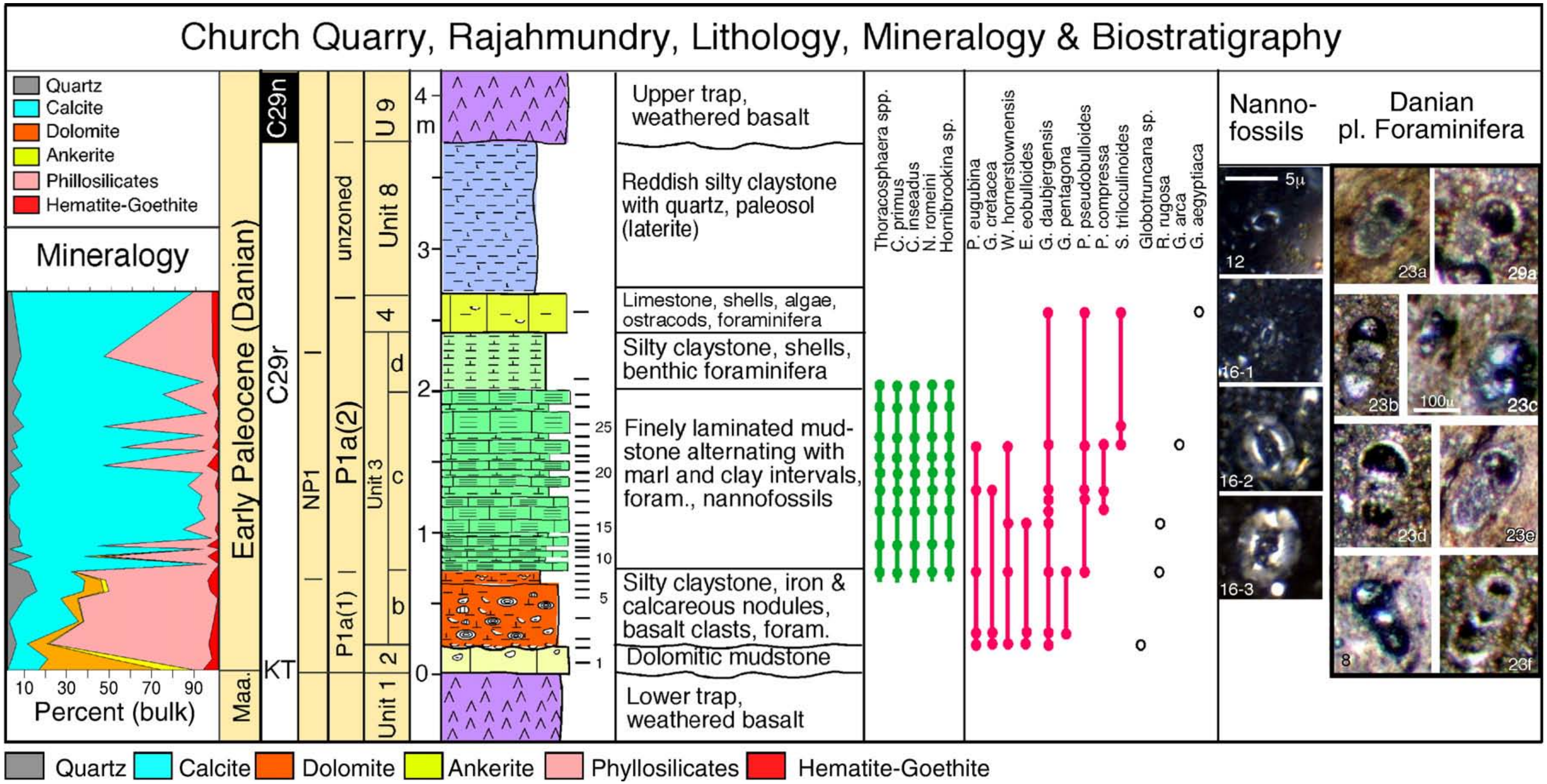

Fig. 4. Church Quarry intertrappean sediments span the earliest Danian zone P1a, similar to Balaji and Government, but the sequence is condensed due to erosion and/or non-deposition as evident by the reduced unit 2 and missing units 5–7. Lithology and bulk rock compositions reflect shallow estuarine to lacustrine or subaerial environments. Photomicrographs of species (numbers keyed to samples in lithologic column), Calcareous nannofossils (scale bar = 5 µ): 12, *Neobiscutum romeinii*; 16-1, *Cruciplacolithus*? cf *C. primus* (small morphotype); 16-2, *Cruciplacolithus* cf. *C. inseadus*; 16-3, *C. primus*. Planktic foraminifera (scale bar = 100 µ): 8, *Parvularugoglobigerina eugubina*? 23a, 29a, *Globoconusa daubjergensis*; 23b, c, e, f, *Parasubbotina pseudobulloides*; 23d, *Subbotina triloculinoides*.

limestone of the latest Maastrichtian age below the lower trap and a Danian (P1b to P2) age for the intertrappean beds (Jaiprakash et al., 1993; Raju et al., 1996), interrupted by frequent hiatuses (Raju et al., 1994). Similarly, the Narasapur well yielded an early Paleocene (Danian) age based on planktic foraminifera (Govindan, 1981; Jaiprakash et al., 1993; Raju et al., 1995, 1996), calcareous nannoplankton (Saxena, and Misra, 1994), dinoflagellates (Mehrotra and Sargeant, 1987) and palynology (Prasad and Pundeer, 2002). Despite all these studies, a firm basis for an early Danian age has remained elusive. Nevertheless, these studies suggest that given closely spaced samples, higher resolution age control may be possible. With this objective in mind we concentrated on the intertrappean sediments between the two Rajahmundry traps in four quarries (Fig. 1). The Rajahmundry area is ideal because the two traps represent part of the Deccan volcanic acme and the geographically longest lava flows that have been tentatively correlated to the Ambenali and Mahalabeshwar Formations (Chenet et al., 2007, in press; Saunders et al., 2007; Jay and Widdowson, 2008).

Here we report on the biostratigraphic age and sedimentary environment of the Rajahmundry intertrappean sediments in four quarries of the KG Basin (Balaji, Government, Church and Duddukuru) based on planktic foraminifera, calcareous nannofossils and sedimentology. The main objectives include: (1) determine the biostratigraphic age of the sediments between the two Rajahmundry traps, (2) determine the position of the K–T boundary relative to the two traps, (3) compare the biostratigraphic age with published magnetic polarity data and radioisotopic ages to obtain better age control, (4) determine the depositional environment based on microfacies analysis, microfaunas and microfloras, and (5) evaluate the timing and biotic consequences of Deccan flood basalt eruptions with respect to the K–T mass extinction.

2. Methods

In each of the four quarries studied for this report the exposed sediments, consisting of limestones, claystones, laterites and lava flows, were examined for macrofossils, lithological changes, unconformities and hardgrounds. The sections were described, measured and sampled for analyses. In the laboratory thin sections were made for lithofacies and microfossil analyses. Washed residues of the silt and clay samples yielded mostly benthic foraminifera in some layers and only rare planktic foraminifera. The reason appears to be environmental (shallow estuarine), rather than preservational, as benthic foraminifera are generally well preserved and rare aragonitic tests of bivalves were observed. Overall carbonate preservation is poor as most claystone and limestone layers are often dolomitic or recrystallized, obliterating the tiny early Danian species. In addition, poor preservation tends to eliminate fragile species in washed residues, though they remain preserved in thin sections. For these reasons, biostratigraphic analysis was based mainly on thin sections where sufficient species could be identified for early Danian biozone determinations. For each sample, thin sections were systematically scanned for foraminifera in the matrix and clasts and the species photographed for illustration and as permanent records. For calcareous nannofossil analysis samples were prepared based on standard methods described in Gardin (2002). Assemblages are moderately to well preserved, low in diversity and rare. Bulk and clay mineral analyses were based on XRD (SCINTAG XRD 2000 Diffractometer), following the procedures of Kübler (1987) and Adatte et al. (1996).

3. Lithology and microfacies

The four Rajahmundry quarry sections studied span the intertrappean sediments between the lower and upper Deccan trap flows (Figs. 2–5). In all four sections, the lithologies are similar, varying only in the thickness of lithofacies and the variable extent of erosion leading to elimination of some lithofacies. Lithologies range from claystone, siltstone, limestones and dolomitic limestones to paleosols (calcrete, laterite), which can be subdivided into 9 different units and 11 microfacies, including the lower and upper traps. This subdivision of lithologic units facilitates correlation of the outcrops and the microfacies aid interpretation of the depositional environment. In the Balaji and Government quarries, all nine lithological units are present (Figs. 2 and 3). In the Church quarry the intertrappean sediments are more condensed for most units, though highly expanded for unit 3 (Fig. 4), whereas at Duddukuru unit 5 is expanded (Fig. 5).

Bulk rock mineralogy was analyzed in three sections. At the Government quarry the bulk rock composition is dominated by phyllosilicates (0.5–98%) and calcite (0–100%), variable dolomite (0–58%) and ankerite (Fe-rich dolomite, 0–33%) and minor quartz (0–17%), hematite and goethite (0–5%, Fig. 2). Similar patterns are observed in the Balaji, Church and Duddukuru quarries (Figs. 3–5). This composition reflects the predominant limestone and silty claystone lithologies.

Dolomite and ankerite are enriched in unit 2 and subunit 3a (Figs. 2–4). These minerals are probably of secondary origin, resulting from Mg and Fe enriched fluids circulating from the underlying basalt. In unit 3b, these minerals are linked to the abundant reworked mudstone clasts eroded from unit 2. Calcite is dominant in units 4 and 6. Phyllosilicates are dominant in the silty claystones (units 3, 5, 7, 8) and linked to paleosols. XRD analysis of the clay fraction reveals that the phyllosilicates are exclusively composed of smectite (an alteration product of basalt) with peak abundance coincident with highest quartz content. High hematite and goethite contents coincide with abundant iron nodules (unit 3, base unit 4).

The distinct lithologies and microfacies of the four sections studied span a distance of less than 4 km and can be easily correlated (Fig. 6). The differential erosion and depositional patterns and variable hydrodynamic conditions reveal the shallow estuarine to lacustrine or subaerial environments, where uneven topography and canyon infilling likely account for the variable thickness or absence of some units (e.g. expanded unit 3, absence of units 5–7, Fig. 6). The lithologic units and eleven microfacies identified (MF 0–10, Fig. 7) permit reconstruction of the paleoenvironments and sea-level fluctuations.

Units 1 and 9 mark the lower and upper Rajahmundry traps, respectively. The lower trap consists of 3–4 lava flows, similar

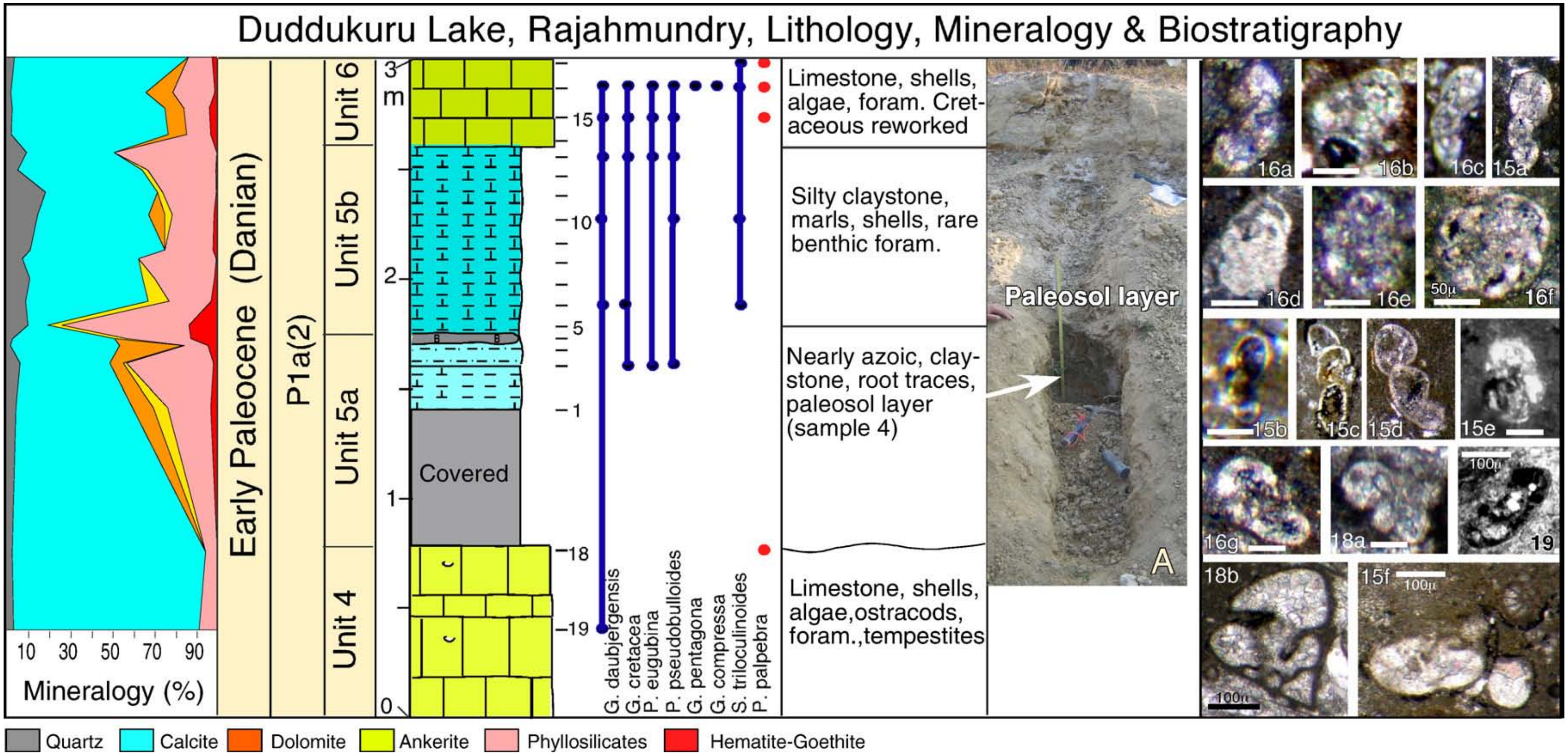

Fig. 5. Dudukuru Lake intertrappean sediments span the earliest Danian zone P1a, similar to the other outcrops, but unit 5 which includes a paleosoil (A) is greatly expanded. Bulk rock mineralogy reflects the shallow estuarine (>calcite) and subaerial (>phyllosilicates) environments. Photomicrographs of species (numbers keyed to samples in lithologic column), scale bar = 50 μ, unless otherwise indicated. 16a. *Globigerina pentagona*; 16b. *Globoconusa daubjergensis*, 16c, 15a, c–e. *Parasubbotina pseudobulloides*; 16d–f, 15b–c. *Parvularugoglobigeria eugubina*; 16 g. *Subbotina triloculinoides*; 18a, unidentified species; 15f, 19. benthic foraminifera; 18b. *Pseudoguembelina palpebra*, reworked Cretaceous species.

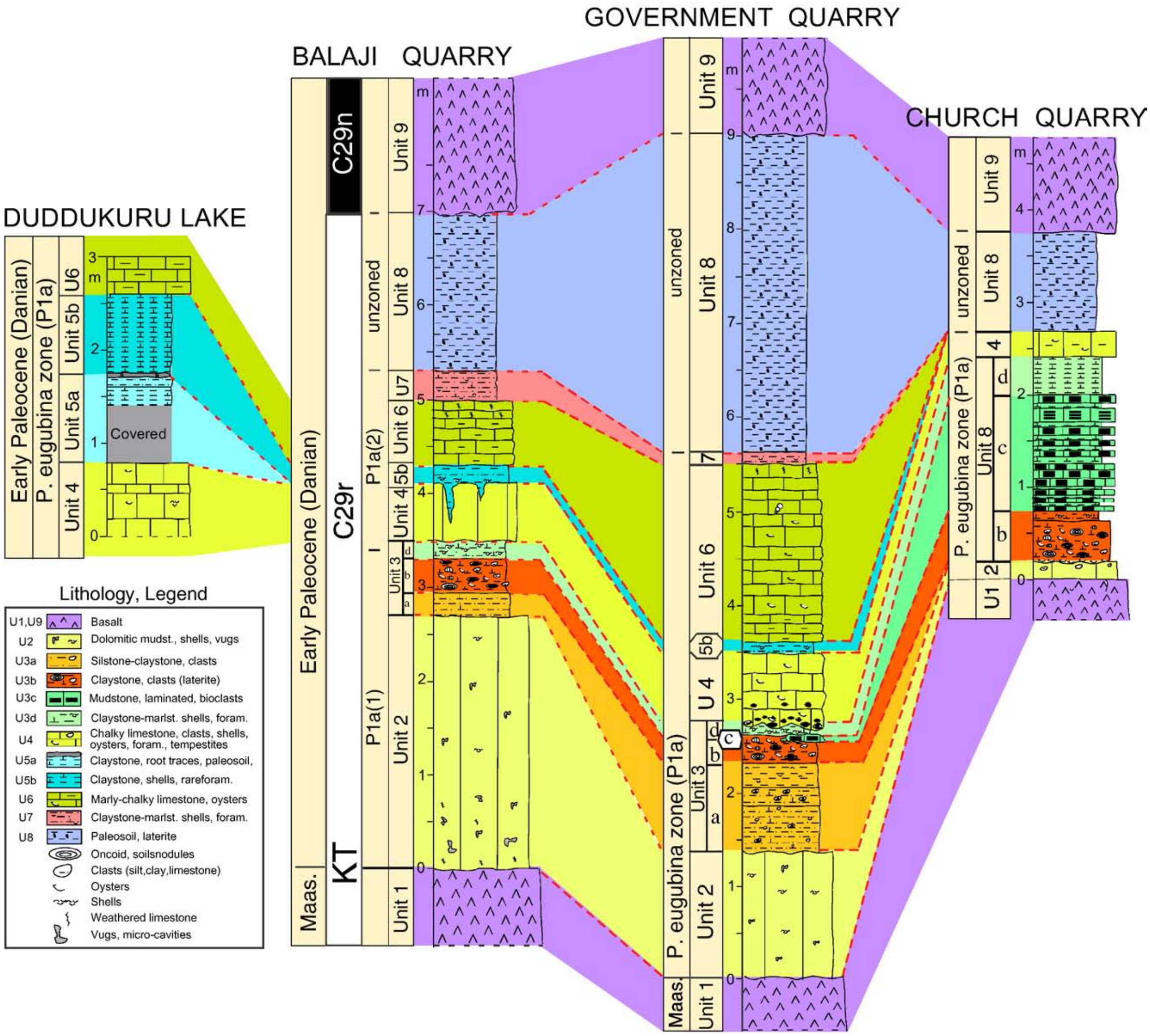

Fig. 6. Lithologic and biostratigraphic correlation of the four outcrops studied. The variable thicknesses and occasional absences of lithological units reflect the paleotopography, incised valleys, variable rates of erosion due to current activity, location within the estuarine environment and sea-level fluctuations.

to the Palakollu-A well, though they lack significant intertrappean sediments (Knight et al., 2003, 2005) (Fig. 8). In the Rajahmundry area, the top unit-a of the lower trap consists of subhorizontally layered basalt (Figs. 8, 1a and 2a). The surface of unit-a is strongly weathered with holocrystalline textures and microcavities, which are filled by later stage minerals (Fig. 7, MF 0). Unit-b is a columnar basalt (Fig. 8, 1b and 2b). The basal lava unit-c consists of pillow lava-like structures, which suggest underwater deposition. Geochemically precipitated calcareous sediments are embedded between the pillow lavas (Figs. 8, 3c and 4c). These units correlate with the three major lava flows in the Palakollu-A well, which were deposited near the end of the Maastrichtian (Jaiprakash et al., 1993).

Intertrappean unit 2 is up to 2.5 m thick in the Government and Bajali quarries and reduced by erosion in the Church quarry (Figs. 2–4). This unit consists of dolomitic mudstone with basalt and claystone clasts, roots and microcavities showing geopetal structures partially filled with vadose silt and successive generations of calcite cement. (Fig. 7, MF 3). These features suggest supratidal conditions and significant hydrothermal activity during deposition. The uppermost contact is erosive and corresponds to a calcrete breccia of continental origin (Fig. 7, MF 1) with claystone clasts containing small early Danian foraminifera eroded from an earlier marine unit. Several cementation phases of sparry calcite, including drusy mosaic to blocky «dogtooth cement», suggest a meteoric subaerial depositional environment.

Unit 3 consists of claystones with numerous carbonate clasts. This unit is best developed in the Balaji, Church and Government quarries where it spans from 0.7 to 1.1 m and can be divided into four subunits. The lower part of subunit 3a is characterized by a nearly azoic claystone (Fig. 7, MF 2) with root traces, which may be enriched in fine-grained quartz. Rare bivalves and ostracods occur near the top (Figs. 2 and 3). This microfacies indicates supratidal or floodplain environments. The upper part of subunit 3a corresponds to shallow, restricted, estuarine conditions (MF 4)

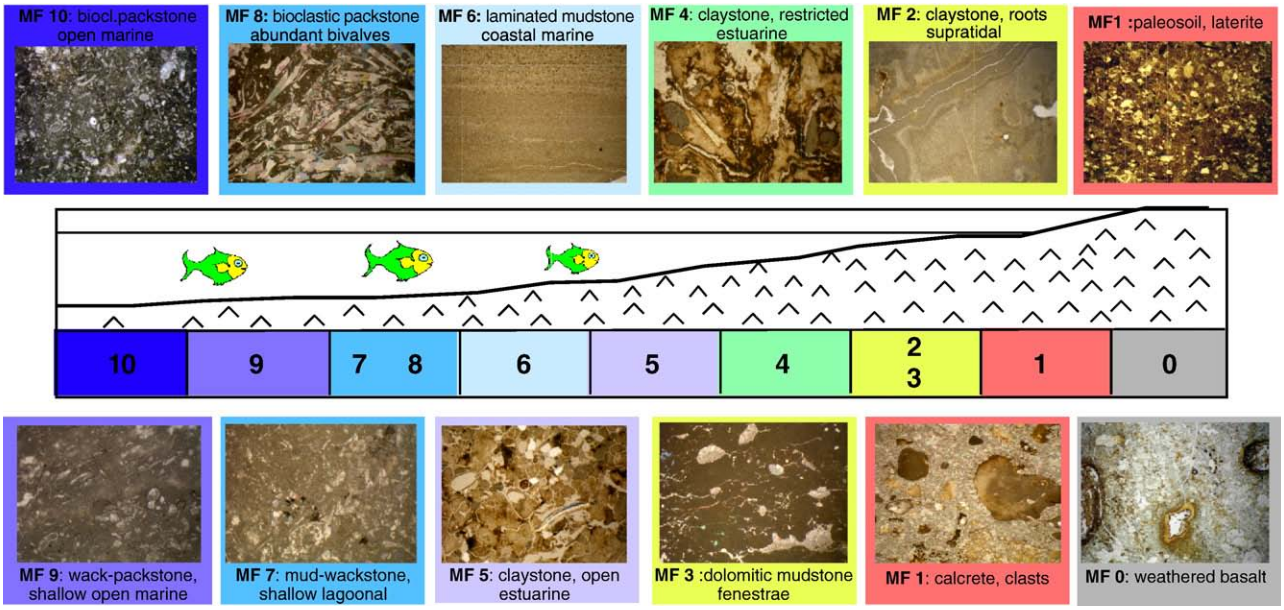

Fig. 7. Microfacies MF 0 to MF 10 and their depositional settings within the shallow estuarine to subaerial environments. See text (Section 3) for details.

marked by silty claystone with rare benthic foraminifera and oysters, dolomitic mudstone and claystone clasts with small early Danian planktic foraminifera (Figs. 2–4).

Subunit 3b is well developed in all three quarries and composed of 40–70 cm thick red to brownish claystone with very abundant calcareous and iron oxide (hematite) nodules reworked from paleosols, and large basalt clasts. The presence of abundant bivalves fragments (mainly oysters) and rare benthic and planktic foraminifera indicates marine–estuarine conditions (Fig. 7, MF 5). The higher abundance of shells and smaller clast size, as compared with MF 4, reflects more open estuarine conditions.

Subunit 3c consists of a finely laminated calcareous mudstone with rare bioclasts of marine origin. This interval is 1.2 m thick at the Church quarry, but forms only a 10 cm thick lens at the Government quarry and is absent at Balaji (Figs. 2–4 and 6). The laminations, which consist of alternating fine carbonate layers containing sparse coccoliths and slightly

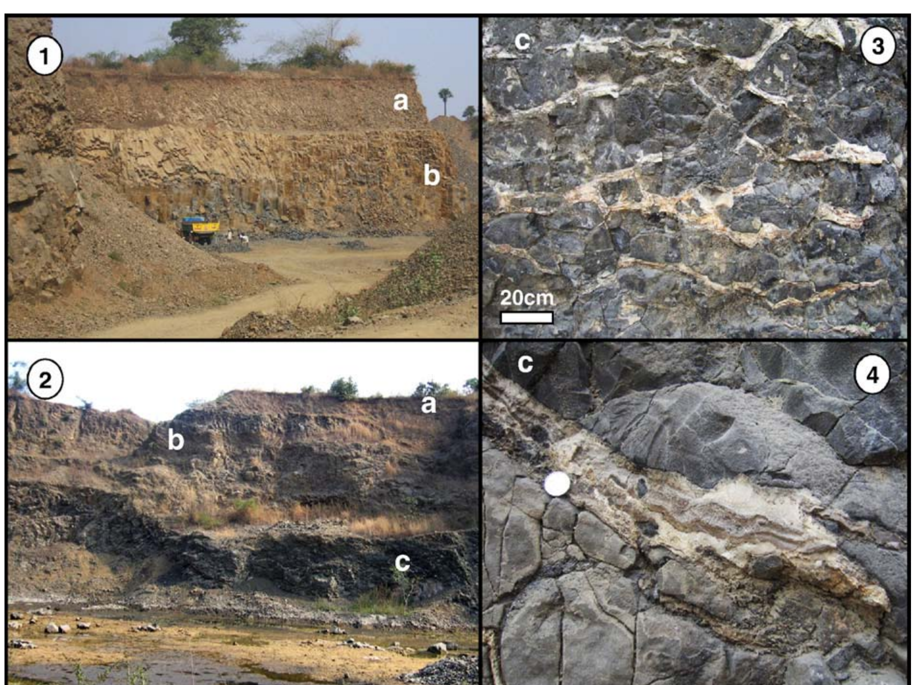

Fig. 8. Three major lava flows of the lower trap are exposed in quarries of Rajahmundry. 1. Uppermost (a) and middle (b) lava flows. 2. Upper (a), middle (b) and lower (c) lava flows. 3. Intratrappean sediments are thin, discontinuous, recrystallized and devoid of microfossils. 4. Pillow-like lava structures of unit c indicate that some lava flows erupted under water.

coarser layers enriched in continental organic matter, suggest seasonal cyclicity in a low energy environment (Fig. 7, MF 6). No bioturbation is observed. The uppermost subunit 3d is 10–20 cm thick and present in all three sections. This greenish claystone contains rare benthic and planktic foraminifera, nannofossils, shell fragments, fossil fishes and rare small calcareous and iron oxide nodules (Fig. 7, MF 4–5).

Unit 4 consists of a 0.2 to 0.8 m thick limestone, which is present in all four localities examined (Fig. 6). The base of this unit is erosive, graded and contains abundant clasts reworked from the underlying layer, as well as from older Cretaceous lithologies. This indicates deposition in a shallow but open marine environment with occasionally high-energy hydrodynamic conditions. Several discrete layers show hummocky stratification and are significantly enriched in bivalve shells and fish bones, suggesting tempestites (Fig. 7, MF 8). The interval above is characterized by a mudstone–wackestone microfacies with common bioclasts, including shells and fish bone fragments, gastropods, algae, benthic and planktic foraminifera and ostracods (Fig. 7, MF 7). This corresponds to a shallow lagoonal environment. The upper part of unit 4 reflects a more open marine environment, as indicated by Danian planktic foraminifera, and rare algae and benthic foraminifera (Fig. 7 MF 9). At Balaji, the top of unit 4 is marked by deeply incised karst features (Fig. 3), which are infilled by sediments from the overlying claystone unit 5 (Fig. 7, MF 4).

Unit 5 is a 10–20 cm thick silty–claystone at the Balaji and Government quarries, 1.2 m thick at Duddukuru and absent at the Church quarry (Fig. 6). At Duddukuru unit 5 can be subdivided into subunits 5a and 5b based on a well-developed paleosol horizon (Fig. 5). Subunit 5a consists of a nearly azoic clay layer with root traces, rare shell fragments and foraminifera (Fig. 7, MF 2). A paleosol horizon at the top of subunit 5a (MF 1) probably corresponds to the karst contact observed at the base of unit 5b at Bajali (Fig. 3). Subunit 5b, which is present in all sections, except the Church quarry, and consists of a silty claystone with rare benthic and planktic foraminifera, common shells and fossil fishes (e.g., *Eotrigonodon* sp., *Chrysophrys* sp. and *Pycnodus* sp., MF 4). Unit 5 thus reflects supratidal to terrestrial environments followed by a return to restricted estuarine conditions.

Unit 6 is the most calcareous lithology, consisting of progressively thinner limestone beds ranging from 0.7 to 1.8 m in thickness in three of the four quarries examined (Fig. 6). The lower part of unit 6 consists mainly of wackstones (MF 7) occasionally enriched in shell fragments and reworked Cretaceous planktic foraminifera (Fig. 7, MF 8), and characterized by sigmoidal cross-bedding (megaripples) suggesting tempestites (similar to unit 4). The upper 50 cm consists of a finer limestone, or packstone with common Danian planktic foraminifera, few to common reworked Cretaceous species and lesser benthic fauna (Figs. 2 and 7, MF 10). This reflects a shallow but more open marine environment than MF 9.

An erosive surface marks the contact between the limestone of unit 6 and silty clay of the overlying unit 7, which consists of a 10–30 cm thick silty claystones with rare shell fragments and foraminifera (Figs. 2, 3 and 7, Mf 4). Unit 8 consists of a prominent 1 m to 3.5 m thick red paleosol enriched in laterite-derived material and fine-grained quartz (Figs. 2–4 and 7, MF 1). This represents shallowing from shallow estuarine to terrestrial environments. The upper trap (unit 9) was deposited in a terrestrial environment and the basalt is strongly weathered, similar to the lower trap.

4. Biostratigraphy

4.1. Planktic foraminifera

The K–T boundary is easily identified worldwide in planktic foraminifera by the mass extinction of all tropical–subtropical Cretaceous species (2/3 of the assemblages), the immediate increased abundance (up to 90%) of the disaster opportunist survivor *Guembelitria cretacea* (Keller et al., 1995, Keller, 2002, 2002, 2003; Keller and Pardo, 2004), and the evolutionary first appearances of Danian species (e.g., *Parvularugoglobigerina extensa*, *Woodringina hornerstownensis*, *Globoconusa daubjergensis*, *Eoglobigerina eobulloides*, Fig. 9). The interval from the first appearance of Danian species to the first appearance of *P. eugubina* and/or *P. longiapertura* generally marks the boundary clay zone P0, which usually is enriched in iridium. The total range of *P. eugubina* marks biozone P1a (MacLeod and Keller, 1991, 1994; Keller et al., 1995; Koutsoukos, 1996; Luciani, 2002; Molina et al., 2005). Within this range, the first appearances of *Parasubbotina pseudobulloides* and *Subbotina triloculinoides* subdivide biozone P1a into subzones P1a(1) and P1a(2) (Fig. 9) (Keller et al., 1995). The first Danian nannofossil biozone NP1 marks this early Danian interval.

Planktic foraminifera in sediments of the Government, Balaji, Church and Duddukuru sections are rare due to the shallow marine environments. Nevertheless, good age control can be obtained from assemblages in marine facies, such as limestone, claystone and clay and mudstone clasts. The first Danian planktic foraminifera are found in claystone clasts of unit 2, which overlies the lower Rajahmundry trap. In the Government quarry, these clasts contain tiny early Danian species, including *Parvularugoglobigerina eugubina*, *G. daubjergensis* and the Cretaceous survivor and disaster opportunist *G. cretacea* (Fig. 10). The same early Danian species are also present in claystone clasts of unit 2 in the Balaji quarry (Fig. 3). Two additional species, *E. eobulloides* and *W. hornerstownensis*, are present in clasts from unit 2 of the Church Quarry (Fig. 4). Unit 2 is thus younger than zone P0 and of early zone P1a age, as also indicated by the presence of the index species *P. eugubina* in the overlying units. The claystone clasts with early Danian species indicate erosion of a claystone layer that could have been deposited after the arrival of the topmost lava flow of the lower Rajahmundry trap.

Claystones of units 3, 3b and 3c in the Balaji, Church and Government quarries, contain more diverse early Danian biozone P1a(2) assemblages, including *P. eugubina*, *G. daubjergensis*, *P. pseudobulloides*, *Chiloguembelina crinita*, *W. hornerstownensis*, *S. triloculinoides* and *Globigerina* (*Eoglobigerina*) *pentagona*, (Figs. 3–5 and 10). Such diverse early Danian assemblages first appear about 100 ky after the K–T mass extinction (MacLeod and Keller, 1991; Keller et al., 1995). Rare reworked late

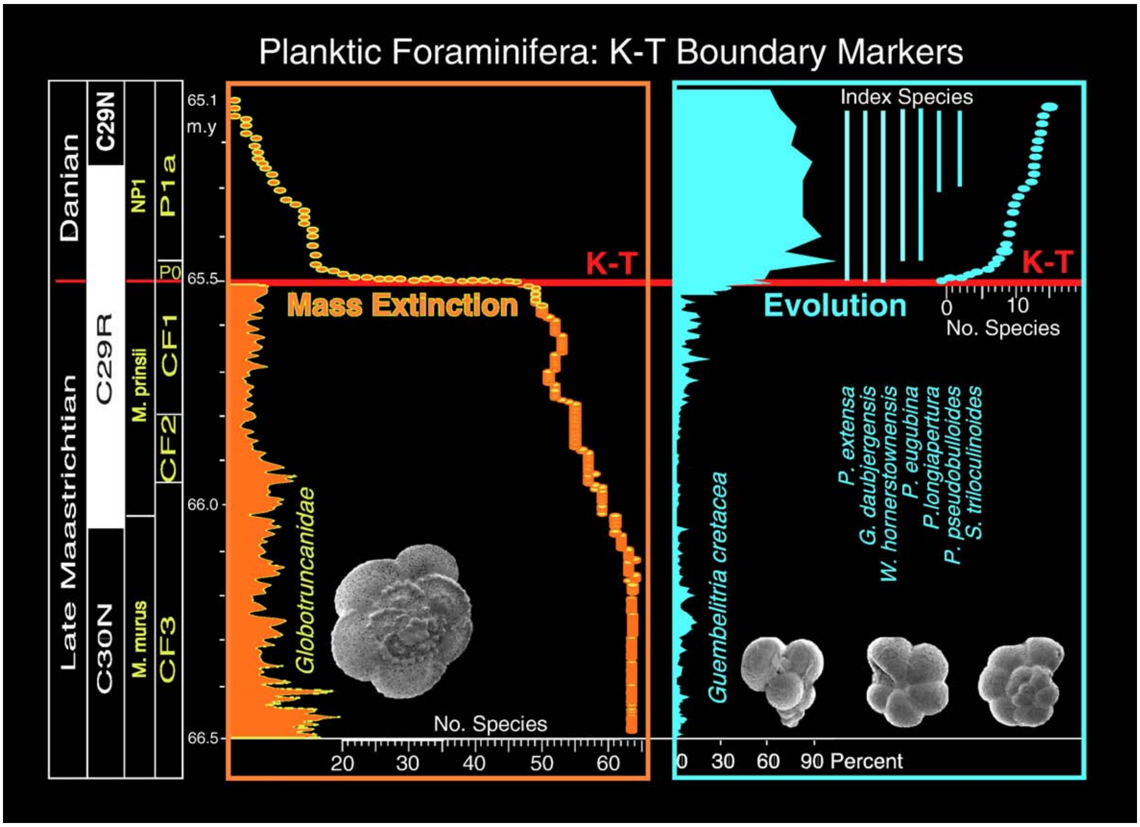

Fig. 9. K–T boundary markers in planktic foraminifera from the El Kef, Tunisia, stratotype section. The KTB is easily identified based on the mass extinction of planktic foraminifera, sudden dominance of the disaster opportunist *Guembelitria cretacea*, and the first appearances of Danian species in zone P0 and P1a.

Maastrichtian species are present. Unit 3 is not exposed at Duddukuru.

Benthic foraminifera in the upper part of unit 3 of the Government quarry are well preserved, nearly monospecific and dominated by nonionids (96–97%, size fraction >125 µm). In this assemblage, *Nonion kingi* (73–44%) and *Protelphidium adamsi* (23–46%) dominate, whereas *Protoelphidium duddukuruense*, *Discorbis toddae*, *Fissurina levigata* and *Rosalina* sp. are minor components totaling <4% of the assemblage. The very low benthic species diversity reflects the high biotic stress of littoral environments.

In unit 4 of the Church quarry rare *G. daubjergensis*, *S. triloculinoides* and *P. pseudobulloides* were recognized (Fig. 4) and *G. daubjergensis* at Duddukuru (Fig. 5). Oysters are common. In unit 5 of the Government quarry, a diverse early Danian subzone P1a(2) assemblage is present (Fig. 10). Benthic foraminifera in this interval are very similar to those in unit 3 and dominated by *N. kingi* (77%) and *Protoelphidium adamsi* (22%). At Duddukuru, the claystone unit 5 is expanded to more than 2 m. In subunits 5a and 5b, *P. eugubina*, *G. cretacea*, and *P. pseudobulloides* mark subzone P1a(2). Algae, shallow-water benthic foraminifera and clasts with rare reworked Cretaceous foraminifera are also present (Fig. 5).

In the Government quarry and Duddukuru Lake, the limestone unit 6 is partly recrystallized, but contains diverse subzone P1a(2) assemblages, including *P. eugubina*, *G. daubjergensis*, *G. pentagona*, *Globanomalina compressa* and *S. triloculinoides* (Figs. 5 and 10). This limestone also contains common reworked late Maastrichtian planktic foraminifera, particularly near the base of unit 6 (Fig. 10), including *Globotruncana arca*, *G. dupeublei*, *G. rosetta*, *Contusotruncana contusa*, *R. plicata*, *Rugoglobigerina rotundata*, *R. rugosa*, *R. macrocephala*, *Pseudoguembelina palpebra*, *Laeviheterohelix glabrans* and *Globotruncanella petaloidea*. Reworked Cretaceous species are also present in this limestone in the Balaji quarry and Duddukuru.

The origin of this late Maastrichtian assemblage can be inferred from the Palakollu-A well, located about 50 km east of the Rajahmundry quarries. In this well, the lower Rajahmundry trap consists of four lava flows separated by intertrappean sediments devoid of planktic foraminifera (Jaiprakash et al., 1993). Only in the limestones below the lowermost lava flow are similar late Maastrichtian assemblages observed. This suggests that the reworked species in unit 6 probably derived from erosion of late Maastrichtian limestones that were deposited before the arrival of the lower trap lava flows. Regional uplift associated with the volcanic activity may have resulted in erosion of late Maastrichtian sediments, particularly in areas not covered by the Rajahmundry trap flows.

Erosion and/or non-deposition of units 6 and 7 in the Church quarry and a significantly reduced limestone deposition at Balaji, as compared with the Government quarry (Fig. 6) indicate high current activity in subzone P1a(2). This is evident by the variable thickness of unit 7 (e.g. 30 cm thick at Balaji,

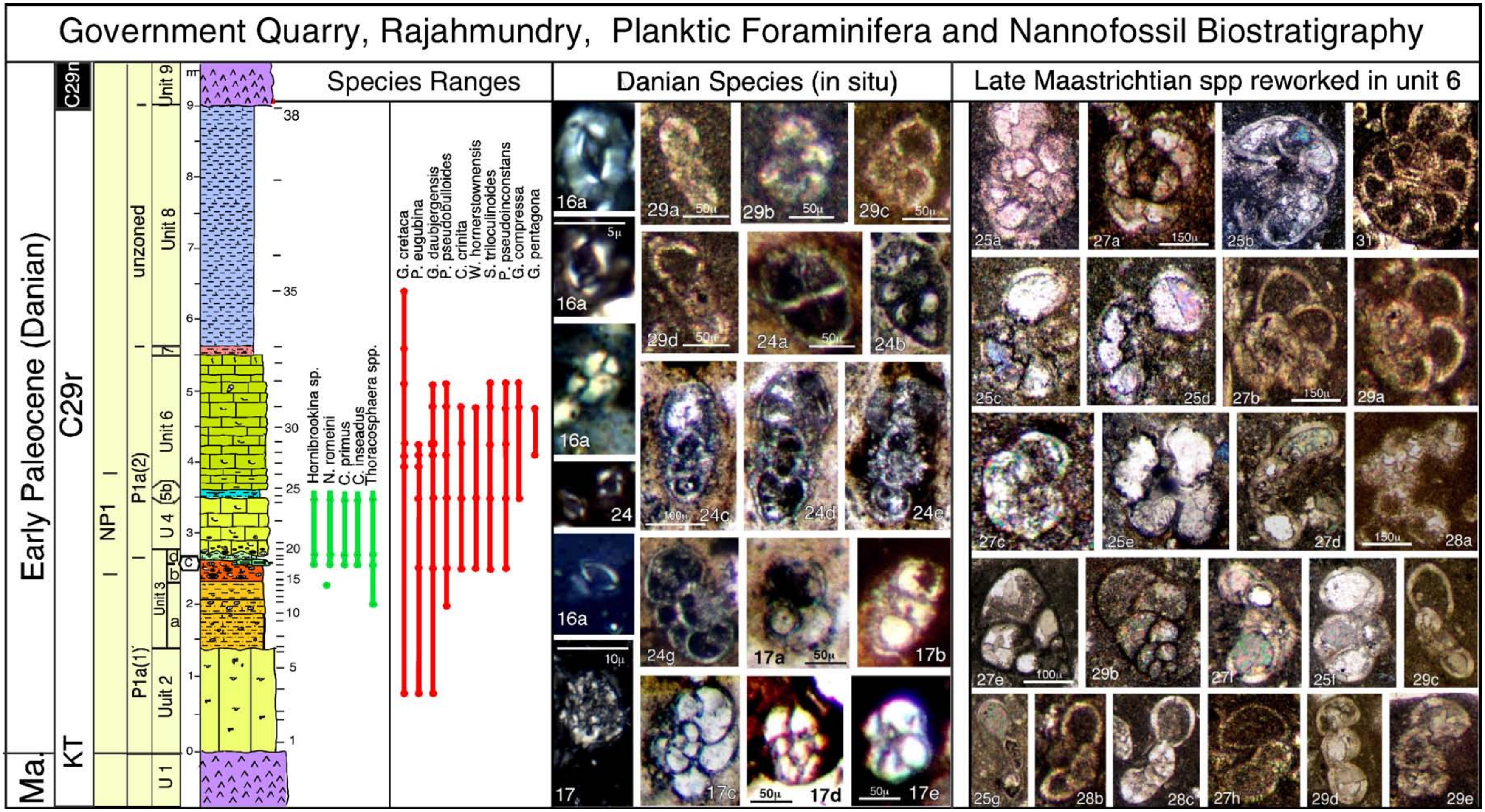

Fig. 10. Government quarry, biostratigraphy of intertrappean sediments. Photomicrographs of species (numbers keyed to samples in lithologic column): Calcareous nannofossils (scale bar = 5 µ for the first five species): 16a-1, *Cruciplacolithus* cf. *C. inseadus*; 16a-2, *Cruciplacolithus primus*; 16a-3, *Octolithus multiplus*; 24, *Neobiscutum romeinii*; 16a-4, *Hornibrookina* sp.; 17, *Thoracosphaera* spp fragment. Danian planktic foraminifera: 29a, b, 17d, e, *Parvularugoglobiernina eugubina*; 29c, 24a, 17a, *Globoconusa daubjergensis*; 24b, *Chiloguembelina sp*; 29d, *Globanomalina compressa*; 24c, g, d, 17c, *Parasubbontina pseudobulloides*; 24e, *Subbotina triloculinoides*. Cretaceous reworked species (scale bar = 150 µm for upper three rows; 100 µm for lower two rows). 25a, c, d, 27b, 29a, *Globotruncana arca*; 25b, 27a, *Contusotruncana contusa*; 31, *Globotruncana dupeublei*; 27c, *Globotruncana rosetta*; 25e, 27d. *Contusotruncana plicata*; 28a. *Rugoglobigerina rotundata*; 27e. *Pseudoguembelina palpebra*; 29b, *Laeviheterohelix glabrans*; 27f, 29e, *Rugoglobigerina rugosa*; 25f, 27h, 29b, *Rugoglobigerina macrocephala*; 29c, 25g, 28b, c, *Globotruncanella petaloidea*.

10 cm at Government and absent at the Church quarry). At the Balaji quarry, a subzone P1a(2) assemblage was identified, including *P. eugubina*, *G. daubjergensis*, *P. pseudobulloides* and *Praemurica compressa* (Fig. 3, unit 7). No biostratigraphic age could be determined for the paleosol unit 8 that underlies the upper Rajahmundry trap.

Planktic foraminifera thus indicate that units 2 to 7 of the intertrappean sediments in the Rajahmundry sections were deposited in the early Danian zone P1a. Early Danian zones P0 and P1a encompass the 200 ky interval of C29r above the K–T boundary, with the extinction of *P. eugubina* coincident with the top of C29r, or very base of C29n (MacLeod and Keller, 1991). Based on the presence of zone P1a, the intertrappean sediments (units 2–7) were thus deposited during C29r above the K–T boundary. Knight et al. (2003, 2005) who collected most of their samples in outcrops near Duddukuru, determined that the upper trap was deposited in C29n. This is in agreement with the biostratigraphic age determined from the intertrappean sediments.

4.2. Calcareous nannofossils

The boundary between the Maastrichtian and Danian is marked by the disappearance of Cretaceous taxa, or the first occurrence of the *Thoracosphaera* acme in low latitudes (Martini, 1971). In the Rajahmundry sections, these characteristics were not observed, indicating that the KTB event was not encountered. It is noteworthy that all species commonly present in the early Danian, including *Thoracosphaera* sp, *Neobiscutum romeini* and *C. primus* were already present, though very rare, in uppermost Maastrichtian sediments in neritic environments (Gardin and Monechi, 1998; Gardin, 2002; Mai et al., 2003). In the early Danian these species thrived and the bulk of the characteristic late Cretaceous species disappeared. It is this assemblage change that permits recognition of the early Danian biozone NP1.

In the Rajahmundry sections, calcareous nannofossils are more sporadic and sparse than planktic foraminifera. In the Government quarry nannofossil assemblages were observed in units 3 and 5 (Fig. 10), which contain *Thoracosphaera* spp. fragments, *N. romeini*, *Crucuplacolithus primus*, *C. inseadus*, *Cruciplacolithus*? sp. and *Hornibrookina* sp. These assemblages are indicative of the early Danian biozone NP1, consistent with the zone P1a age assignment based on planktic foraminifera (Fig. 9).

In the Balaji quarry nannofossils are restricted to units 3 and 5 also, but the assemblages are sparse consisting of a few *Thoracosphaera* spp. fragments and very rare holococcoliths, such as *Octolithus multiplus* (Fig. 3). No coccolith species were encountered. The calcareous dinoflagellate *Thoracosphaera*, as well as holococcoliths *O. multiplus* and *L. duocavus*, are characteristic components of the earliest Danian post K–T assemblages (Mai et al., 2003). Calcareous nannofossils present in the laminated mudstones of unit 3c (Fig. 3) are identical to those observed in the Government quarry and indicative of the early Danian biozone NP1.

The rare occurrence of calcareous nannofossils and the low diversity assemblages are not due to poor or selective preservation, but rather the shallow environmental setting in which few calcareous nannofossil species thrived. The assemblages found in the Rajahmundry intertrappean sediments are almost exclusively composed of *Cruciplacolithus*, including the rarely reported *C. inseadus*, and dwarf coccolith species, such as *N. romeini*. Some modern *Cruciplacolithus* species (i.e., *C. neohelis*) are adapted to coastal, shallow-water environments (Fresnel, 1986) and their Danian ancestors may have inhabited similar high-stress shallow environments (Medlin et al., in press). This interpretation is supported by facies analysis that reveals shallow coastal marine environments.

5. Discussion

5.1. Age of Rajahmundry traps and main Deccan eruptions

5.1.1. Magnetic polarity and $^{40}Ar/^{39}Ar$ age constraints

The lower and upper Rajahmundry Deccan traps are in reversed and normal polarity zones in C29r and C29n, respectively (Vandamme and Courtillot, 1992; Subbarao and Pathak, 1993; Baksi, 2005). The best age determinations to date are based on ^{40}K/^{40}Ar (absolute) and ^{40}Ar/^{39}Ar (relative) dates of plagioclase separates (Fig. 10) (Chenet et al., 2007; Knight et al., 2003, 2005; Baksi, 2005). Error bars for these radiometric ages are large (1% or 0.6 m.y.), which permits no determination of the KTB position. Nevertheless, radiometric ages for the upper trap are well within C29n, but the age for the lower trap is not as well constrained, though still overlaps with C29r of Cande and Kent (1991).

The main Deccan eruptions are in C29r and encompass 2500 m to 3500 m that account for 80% of the total Deccan volume extruded (Chenet et al., 2007, in press; Saunders et al., 2007; Jay and Widdowson, 2008). Four samples from the top of this sequence, the topmost Ambenali and Mahabaleshwar Formations, yielded a mean ^{40}K/^{40}Ar age of plagioclase separates of 64.5 ± 0.6 Ma (Chenet et al. 2007). This age straddles the C29r/C29n reversal boundary and is in good agreement with ages derived from the upper Rajahmundry trap (Fig. 11). Three samples from the lowermost Jawhar Fm. yielded a mean age of 64.8 ± 0.6 Ma. These ages are consistent with the ^{40}Ar/^{39}Ar ages derived from the Rajahmundry traps (Knight et al., 2003, 2005; Baksi, 2005). Despite this good agreement, correlating the two traps to specific eruption phases in the main Deccan remains speculative. One might argue that these radiometric data combined with biostratigraphic data suggest that the upper and lower Rajahmundry traps correlate within the Mahalabeshwar and Ambenali Formations, respectively.

5.1.2. Biostratigraphic age constraints

As radiometric dating has steadily improved to error bars of only a few hundred thousand years, and scientists have tried to decipher and date the sequence of critical events in Earth's history with ever greater age control, the *Law of Superposition* has remained more relevant than ever. This is particularly so for the K–T transition, where a sequence of closely spaced, but separate events (i.e., mass extinction, Ir anomaly, Chicxulub impact, Deccan volcanism, sea-level lowstand, climate change) occurred over a few hundred thousand years. But radiometric dating cannot decipher the order of such closely spaced events

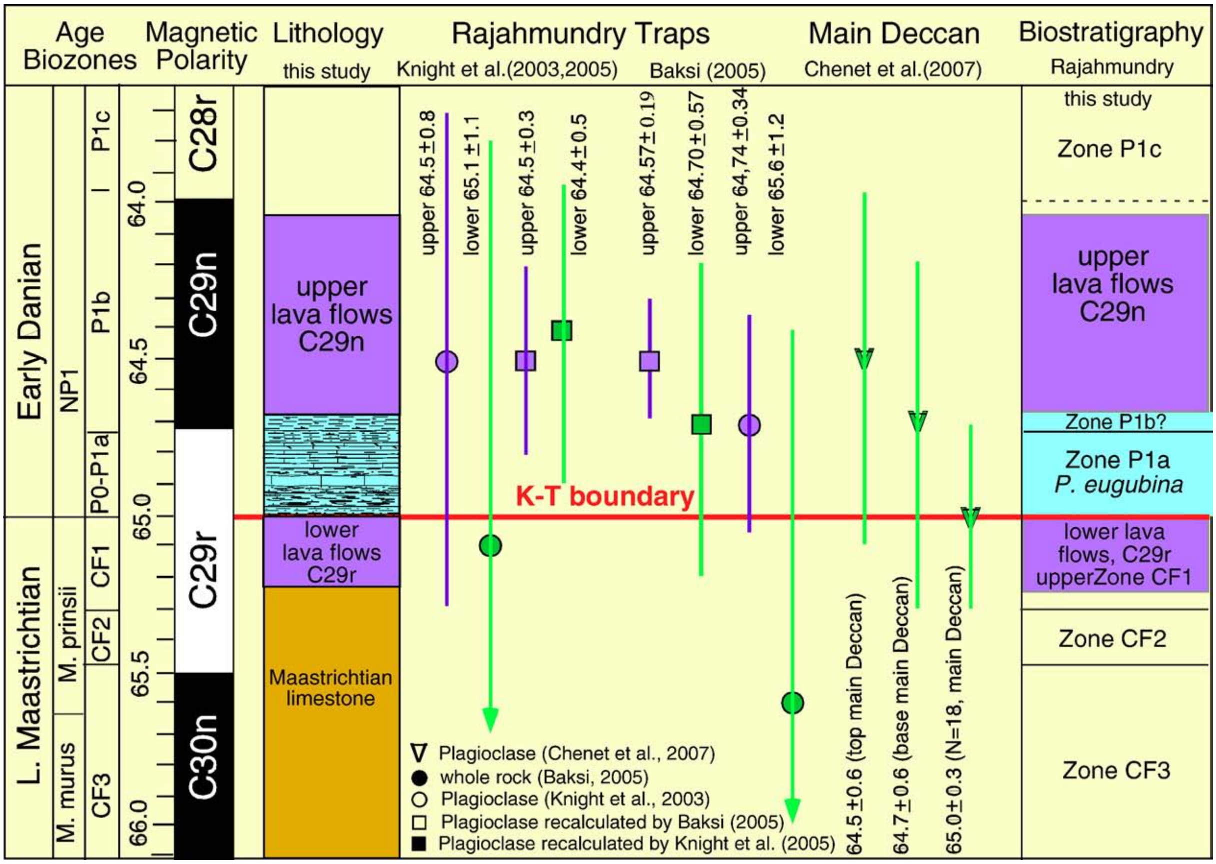

Fig. 11. $^{40}K/^{40}Ar$ and $^{40}Ar/^{39}Ar$ ages of the Rajahmundry Deccan traps (Knight et al., 2003, 2005; Baksi, 2005) and the main Deccan volcanic province (Chenet et al., 2007) yield ages with an accuracy of 1%. Ages for the upper trap are well within C29n, but ages from the lower trap are less well constraint. Planktic foraminiferal biostratigraphy places the lower trap in the latest Maastrichtian with the uppermost lava flow near the K–T mass extinction.

because they fall within the error limits of the dating method, which is typically 1% for the KTB transition. In such cases, the only way to determine the time sequence of events — and hence the potential cause-and-effect — is the stratigraphic relationship of biotic and abiotic events over time, as seen in the sediments and fossils within each rock unit. The power of biostratigraphy as a relative dating tool is clearly demonstrated in dating the intertrappean sediments and thus the Rajahmundry traps.

5.1.3. Age of lower trap flows

In Rajahmundry sediments below the lower trap are of terrestrial origin (Fig. 12-A), but a clue to the basal age can be gained from the Palakollu-A well, located about 50 km east of the Rajahmundry quarries (Raju et al., 1991; Jaiprakash et al., 1993; Raju et al., 1996). In this well, the lower trap consists of 3–4 lava flows separated by intertrappean sediments. Planktic foraminiferal assemblages below these lava flows indicate deposition sometime during the latest Maastrichtian (Jaiprakash et al., 1993). More precise age determination awaits re-analysis of existing wells and new drilling of a continuous sequence through the lower trap, intertrappeans and late Maastrichtian sediments below.

The age at the topmost lava flow of the lower trap can be estimated based on planktic foraminiferal biostratigraphy of the overlying intertrappean sediments. Unit 2, which overlies the lower trap, indicates an early Danian subzone P1a(1) age. The presence of claystone clasts with P1a(1) assemblages in unit 2 indicates that the basal Danian is missing. An estimate of the missing interval can be obtained from biostratigraphy. As noted above, the intertrappean sediments above the lower trap span the interval of C29r above the KTB, or about 200 ky (Cande and Kent, 1991; MacLeod and Keller, 1991; Berggren et al., 1995), which corresponds to biozones P0 and P1a. P0 spans only a few thousand years and subzone P1a(1) marks about 80–100 ky (Fig. 9). Since part of this record is present, the missing early Danian interval is likely no more than 60 ky. There is thus a gap of about 60 ky between the KTB and the earliest Danian deposits above the lower Rajahmundry trap.

In addition, the weathered basalt and erosion surface of the lower trap indicates a relatively short missing interval. For example, in modern humid tropical environments basalt weathers very rapidly (<100 yr) (Benedetti et al., 2003), although in the Deccan province, weathering probably occurred under drier climatic conditions. This is evident in the high smectite (100%) derived from weathering of basalt, which could have occurred over a much longer time period (10–20 ky) (Chesworth et al., 1983). Based on biostratigraphic and mineralogic estimates, the missing time interval between the top of the lower trap and the early Danian P1a(1) planktic foraminifera is likely less than 80–100 ky.

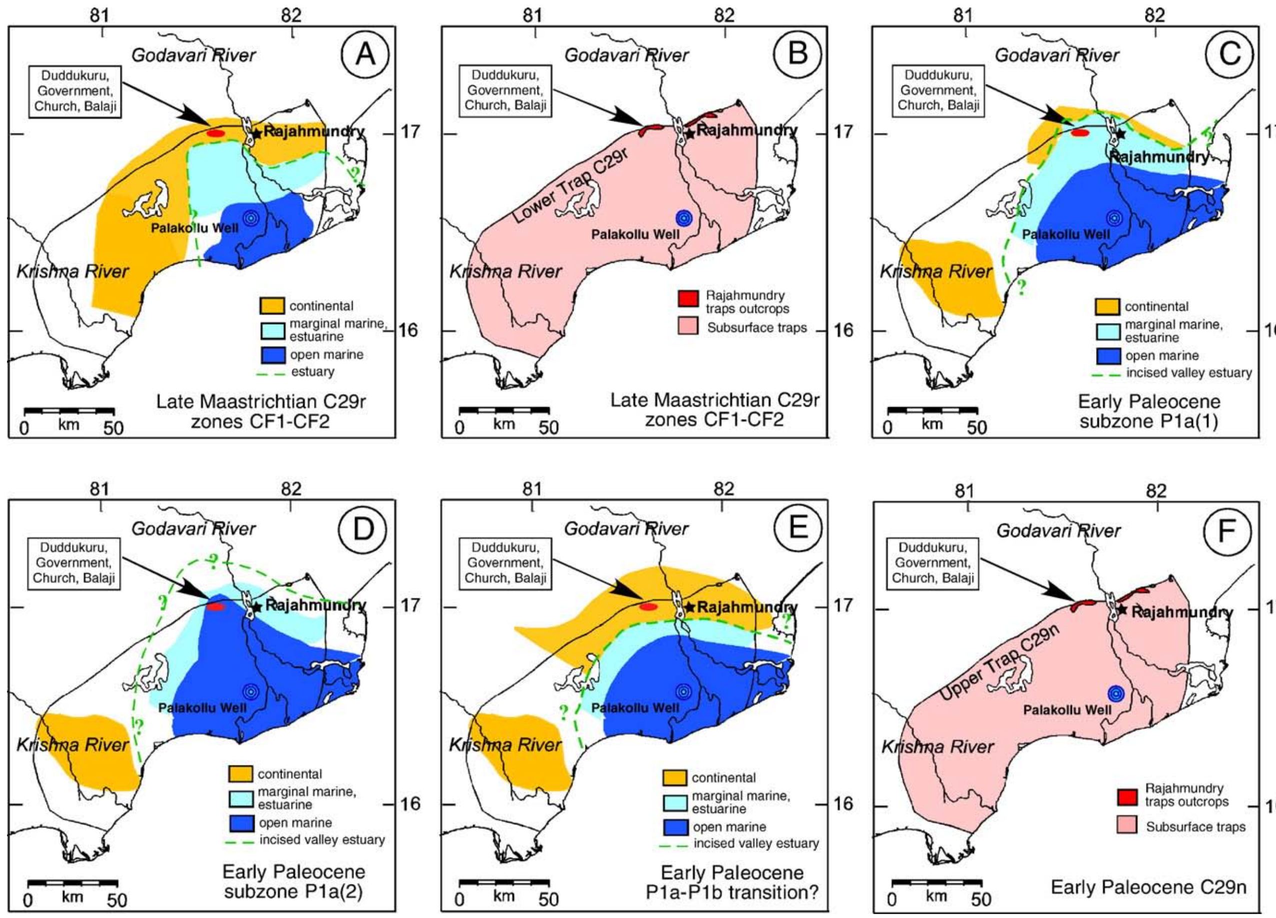

Fig. 12. Paleogeographic maps of the Krishna–Godavari basin for the intertrappean sediments between upper and lower Rajamundry traps illustrate sea-level fluctuations (data from Jaiprakash et al., 1993; Knight et al., 2003). (A) Sea-level regression and emergence during the late Maastrichtian zones CF1–CF2. B: Lower trap emplacement, late Maastrichtian zones CF1–CF2. C: Early Danian lower zone P1a, sea-level rise, marginal marine environment (units 2, 4, 5b) interrupted by periods of emergence (top of unit 2, unit 5a). D: Early Danian upper zone P1a, maximum transgression (unit 6). E: Early Danian sea-level regression during zone P1a–P1b transition (unit 8). F: Early Danian zone C29n, upper trap emplacement.

Recently, $^{187}Os/^{188}Os$ ratios from three deep-sea sections (Atlantic, Pacific and Indian oceans) were used to estimate the timing of the main Deccan phase (Ravizza and Peucker-Ehrenbrink, 2003). $^{187}Os/^{188}Os$ ratios in the global ocean reflect a balance between average riverine input (~1.3) and weathering of mantle derived volcanism and/or extraterrestrial inputs (~0.13). A negative change in the Os ratio therefore reflects increased inputs by volcanism and/or impact(s). The record shows a major decrease in the $^{187}Os/^{188}Os$ ratios (from 0.55 to 0.4) during at least 200 ky (~300–100 ky before the KTB), coeval with the global warming of 4–5 °C commonly attributed to Deccan volcanism (Li and Keller, 1998; Kucera and Malmgren, 1998; Abramovich and Keller, 2003; Wilf et al., 2003). During the last 100 ky of the Maastrichtian Os ratios stabilized at low values (0.4), coincident with global cooling. This was interpreted as marking the end of Deccan volcanism (Ravizza and Peucker-Ehrenbrink, 2003). However, the continued low (0.4) Os values suggest a persistent input of non-radiogenic Os probably from volcanism, similar to the persistent low (0.4) Os values during the first 2 m.y. of the early Paleocene. At the KTB a transient negative Os excursion was interpreted as impact derived input. These Os isotope data thus support the C29r mega-pulse, and do not contradict the biostratigraphic age determination. However, the end of this volcanic mega phase is difficult to date by this method because of the mixed signals of volcanic and impact derived signals.

It is tempting to conclude that the uppermost lava flow of the lower trap (unit-a, Fig. 8) ended at the KTB and that only the earliest Danian sediments were eroded in the sections analyzed. However, the absence of more precise age control for the lower trap lava flows leaves open several other possibilities, including: (1) The lower trap lava flows ended at K–T time and coincided with the mass extinction. (2) The lower trap lava flows ended a short time prior to the KTB. (3) The lower trap lava flows continued into the basal Danian. Among these possibilities the least likely scenario is (3) because reworked clasts are from the early Danian P1a(1). Scenario (1) is questionable because of the weathering and erosion at the top of the lower trap. This leaves (2) as the likely scenario. Although Os isotope data imply that the main phase of Deccan volcanism could have ended as much as 100 ky before the KTB, precise age control is lacking at this time and awaits recovery of marine sequences across the lower trap and Maastrichtian sediments below.

5.1.4. Age of upper trap flows

The age of the upper trap lava flows is not well constrained biostratigraphically because a paleosol (unit 8, Fig. 6) barren of marine microfossils directly underlies it and sediments above the trap flows are of Eocene age (Knight et al., 2003, 2005). However, an early Danian age is indicated by the claystone unit 7 below the paleosol, which contains foraminiferal assemblages characteristic of the upper part of the *P. eugubina* zone P1a. The upper part of zone P1a corresponds to the C29r/C29n transition and is in good agreement with $^{40}Ar/^{39}Ar$ ages (Fig. 11). The persistent low Os isotope values observed during the early Paleocene (Ravizza and Peucker-Ehrenbrink, 2003) most likely represent the last major phase of Deccan volcanism.

5.2. Sea-level fluctuations and paleogeography

During the late Maastrichtian, prior to deposition of the lower Rajahmundry trap in the Krishna–Godavari Basin, the sea level receded by about 50 km (Fig. 12-A) (Raju et al., 1994). The Palakollu-A well in this area marks this regression by a sea-level drop from middle to inner shelf depths, or about 80 m (Jaiprakash et al., 1993; Raju et al., 1994). Although this sea-level regression is probably related primarily to uplift associated with Deccan volcanism (Saunders et al., 2007; Jay and Widdowson, 2008), it was likely exacerbated by the concurrent major global regression that culminated at 65.5 Ma (base of zones CF1–CF2 (Donovan et al., 1988; Li et al., 2000). Deposition of the lower traps occurred in a shallow marine environment in the Palakollu-A area, but primarily terrestrial setting in Rajahmundry. However, the presence of pillow lavas (Fig. 8) indicates that some lava flows may have erupted under water.

The sea-level record during deposition of the intertrappean sediments in the Krishna–Godavari Basin has remained unknown to date (Jaiprakash et al., 1993; Raju et al., 1994, 1996). The current study shows a complex history of an overall deepening environment, interrupted by minor regressions expressed by erosion and/or subaerial deposition (Fig. 13). For example, the top of the lower trap at Government, Balaji and Church quarries show weathered (smectite) and fractured surfaces infilled with the overlying dolomitic mudstone (MF3, Fig. 7). This suggests a significant time of subaerial exposure (i.e., 10–20 ky, (Chesworth et al., 1983; Benedetti et al., 2003) and deposition in a terrestrial setting at least near the end of lower trap volcanism (Figs. 12-B and 13-A). The dolomitic mudstone (unit 2) marks a marginal, very shallow marine environment in the early Danian P1a(1) subzone (Fig. 13-B). At the top of the dolomitic mudstone, a calcrete breccia (MF1) indicates subaerial exposure under seasonal arid climate conditions.

Very shallow, restricted estuarine conditions with high detrital input and a seasonal humid climate mark unit 3a, as indicated by rare ostracods, oysters and benthic foraminifera. In unit 3b an abundance of basalt, mudstone and claystone clasts with earliest Danian planktic foraminifera reflects intense reworking and current activity. Increasing clay and quartz contents (Figs. 2–4) indicate high detrital input and a more humid climate, as also suggested by abundant iron nodules eroded from nearby lateritic soils (Fig. 13-C, D). A rising sea led to lagoonal and coastal marine environments with very low hydrodynamic conditions and a seasonally contrasted climate as reflected by the laminated mudstone with rare marine bioclasts, fish remains (units 3c, d, Fig. 4) and chalky limestone (Figs. 12-C and 13-E). A pronounced sea-level drop occurred at the top of unit 4, as indicated by the strongly karstified (implying increasing humidity) surface at Balaji and erosion at the other outcrops (Figs. 3, 6 and 13-F).

A rising sea returned open marine conditions, depositing first silty claystone (unit 5b) with fish remains, followed by limestone increasingly rich in bioclasts and Danian planktic foraminifera (Figs. 12-D and 13-G). A major sea-level fall (top of unit 7), possibly related to uplift associated with the eruptions of the upper trap, is marked by increased detrital influx reflecting more humid conditions and erosion. Unit 8 at the top of the intertrappean sediments consists of a thick paleosol enriched in laterite-derived elements (MF0, Fig. 7), which indicates deposition in a terrestrial environment. The paleosol consists of quartz, hematite and smectite, which typically form in warm climates with seasonal contrasts in humidity (Chamley 1989). The upper lava flows appear to have been deposited in a non-marine environment (Figs. 12-E, F and 13-H, I).

In summary, sea-level fluctuations interpreted from microfacies indicate a gradual but fluctuating deepening during the early Danian punctuated by significant regressions near the P0/P1a and P1a/P1b transitions. These sea-level changes largely reflect local uplift and subsidence related to the impingement and migration of the plume head on the late Cretaceous paleoshoreline. Despite this local tectonic overprint, the global sea-level regressions at the P0/P1a and P1a/P1b transitions appear to be recognizable (Fig. 13).

5.3. Biotic effects of volcanism

A possible cause–effect relationship between mass extinctions and volcanism has largely been inferred to date from their overall correspondence (Wignall, 2001; Keller, 2005) and the potential effects of volatile fluxes on the global environment (Thordarson et al., 1996; Ivanov et al., 1996; Thordarson and Self, 1998, 2003; Self et al., 2006). An assessment of the biotic effects of volcanism comes from microfossils in sediments that also contain the volcanic rocks, such as the intertrappean sediments of this study and the Ninetyeast Ridge DSDP Site 216 (Keller, 2003). At Site 216, the biotic effects of volcanism were catastrophic and mirror the effects generally attributed to an impact at the KTB. Similarly high-stress environments have been documented across the oceans for late Maastrichtian volcanism and the K–T mass extinction (Keller and Pardo, 2004). But quantitative studies documenting the biotic effects of Deccan volcanism in India have yet to materialize because intertrappean sediments are often of terrestrial or very shallow marine origins. Published records indicate that planktic foraminiferal diversity decreased prior to deposition of the lower trap, with most Cretaceous species nearly eliminated in the intertrappean sediments of the lower trap (Jaiprakash et al., 1993), and possibly extinct at or near the top of the lower trap. How much of this biotic stress is the result of rapid shallowing

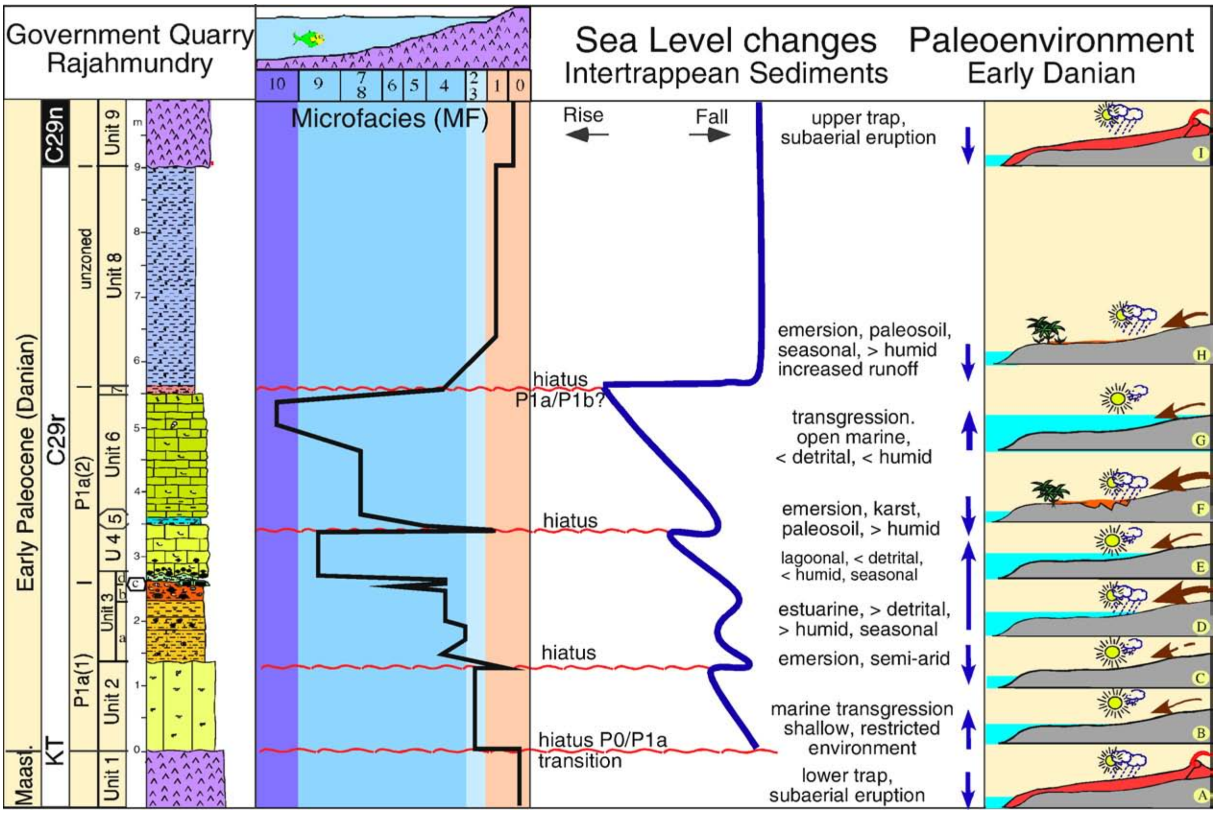

Fig. 13. Sea-level changes and paleoenvironmental interpretation of the depositional settings of intertrappean sediments in the Rajahmundry area based on biostratigraphy and faunal assemblages, lithology, bulk rock mineralogy and microfacies analyses. Sea-level fluctuations likely reflect uplift and subsidence associated with volcanic activity, although low sea levels after emplacement of the lower trap (P0/P1a) and prior to emplacement of the upper trap (P1a/P1b transition) broadly correlate with global regressions. Regressions are marked by emersion and subaerial deposition in humid climates with increased detrital influx. Transgression is marked by estuarine to lagoonal environments with deposition in drier, seasonal climates with decreased detrital influx.

due to uplift related to volcanic activity is unknown at this time. The nature of the K–T biotic catastrophe and the possible cause–effect relationship with the main phase of Deccan volcanism remains to be evaluated from drill cores in the deeper marine environment of the Krishna–Godavari and Cauvery basins.

6. Conclusions

Biostratigraphy and sedimentary facies analyses yield age and environmental information of the intertrappean sediments between the upper and lower Rajahmundry traps of the Krishna–Godavari Basin, NE India.

1. The lower Rajahmundry trap flows in C29r ended very near the K–T boundary.
2. Deposition of the intertrappean sediments occurred in the early Danian zone P1a, which spans C29r above the KTB, or about 200 ky. This is in agreement with K/Ar and Ar/Ar dates, as well as the magnetic polarity data that place the upper Rajahmundry trap in C29n.
3. The lower trap flows are part of the main phase of Deccan volcanism, which occurred during C29r below the KTB, as indicated by biostratigraphic, paleomagnetic and radiometric (K/Ar, Ar/Ar) age constraints.
4. This places the KTB event near the end of the main voluminous Deccan eruptive phase and implies that Deccan volcanism could have been a major contributor to the mass extinction.
5. Sediment deposition occurred in shallow marine environments that fluctuated between supratidal, estuarine, lagoonal and open marine conditions interrupted by periods of subaerial deposition marked by paleosols. Changing sea levels are largely related to uplift and subsidence associated with Deccan volcanism, though eustatic events at the P0/P1a and P1a/P1b transitions appear to be recognizable.

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