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No global-scale transgressive–regressive cycles in the Thanetian (Paleocene): Evidence from interregional correlation

Dmitry A. Ruban [a,*], Svetlana O. Zorina [b], Clinton P. Conrad [c]

[a] Department of Mineralogy and Petrography, Geology and Geography Faculty, Southern Federal University, Zorge Street 40, Rostov-na-Donu, 344090, Russian Federation
[b] Central Research Institute of Geology of Industrial Minerals, Zinin Street 4, Kazan, Republic of Tatarstan, 420097, Russian Federation
[c] Department of Geology and Geophysics, SOEST, University of Hawaii at Manoa, 1680 East-West Road, Honolulu, HI, 96822, USA

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ABSTRACT

A detailed examination of transgressions and regressions that occurred during the Thanetian (58.7–55.8 Ma) may provide an important constraint on the global palaeoenvironment. Seven tectonically "stable" regions (the eastern Russian Platform, Northwestern Europe, Northwestern Africa, Northeastern Africa, the Arabian Platform, the northern Gulf of Mexico, and Southern Australia), represent exceptional records of transgressive–regressive (T–R) cyclicity. Their chronostratigraphic frameworks are sufficiently well constrained to permit accurate correlation. Surprisingly, except for a generally regressive trend occurring in the late Thanetian, no common T–R cycles can be delineated, which indicates an absence of global-scale T–R cyclicity during the Thanetian. Furthermore, we find no clear correspondence between documented T–R patterns and previously reported eustatic changes. We suggest that a warm climate and an absence of major glaciations in the early-middle Thanetian, coupled with only slow eustatic change expected from tectonic processes, stabilized Thanetian eustatic sea level. Regional subsidence or uplift, possibly generated by mantle flow in the form of dynamic topography, governed transgressions and regressions locally and resulted in an inconsistency between T–R cycles in different parts of the globe. The late Thanetian regressive episode, which preceded the Paleocene–Eocene Thermal Maximum, may be linked to an advance of glaciation.

1. Introduction

The Thanetian, the terminal age of the Paleocene epoch, encompassed 2.9 myr, from 58.7 Ma to 55.8 Ma. Chronostratigraphy of this interval is well established (Berggren et al., 1995; Dupuis et al., 2003; Gradstein et al., 2004; Dinarés-Turell et al., 2007; Westerhold et al., 2007, 2008, 2009; Ogg et al., 2008), and permits accurate dating and correlation of geologic events. During the Thanetian, global plate tectonics did not undergo major re-organizations (e.g., Müller et al., 2008), although this relatively "calm" period was followed by the onset of collision between India and Eurasia and intense subduction along the northern Neo-Tethyan margin in the Eocene (Golonka, 2004; Scotese, 2004; Zhu et al., 2005; Müller et al., 2008). Global temperature fell slightly at the beginning of the Thanetian, but then rose significantly toward the Paleocene–Eocene thermal maximum (PETM) (Zachos et al., 2001). Global sea level remained high (Haq et al., 1987; Haq and Al-Qahtani, 2005; Miller et al., 2005; Kominz et al., 2008). Volcanism in the North Atlantic Igneous Province and in the Caribbean region, an extraterrestrial impact, biomass combustion, and voluminous methane release from submarine slope sediments have been proposed as causes for strong, but short-term, environmental perturbations at the Paleocene–Eocene transition (with a peak at ~55 Ma), including an increase in temperature, oceanic anoxia, a re-organization of carbonate platforms, ocean acidification with a consequent change in the carbonate compensation depth, a sea-evel fluctuation, and wide extinctions and turnovers on both land and the sea (Thomas, 1990; Kennett and Stott, 1991; Koch et al., 1992; Kaiho, 1994; Robert and Kennett, 1994; Dickens et al., 1995; Bralower et al., 1997; Hallam and Wignall, 1997; Saunders et al., 1997; Clyde and Gingerch, 1998; Kelly et al., 1998; Hallam and Wignall, 1999; Bains et al., 1999; Katz et al., 1999; Norris and Röhl, 1999; Wignall, 2001; Zachos et al., 2001; Jolley et al., 2002; Speijer and Morsi, 2002; Speijer and Wagner, 2002; Zachos et al., 2003; Kent et al., 2003; Wing et al., 2003; Kahn and Aubry, 2004; Wing, 2004; Zachos et al., 2005; Cramer and Kent, 2005; Thomas and Bralower, 2005; Gingerich, 2006; Higgins and Schrag, 2006; Maclennan and Jones, 2006; Collinson et al., 2007; Takeda and Kaiho, 2007; Smith et al., 2007; Weijers et al., 2007; Moore and Kurtz, 2008; Panchuk et al., 2008; Scheibner and Speijer, 2008; Alegret et al., 2009; Boucsein and Stein, 2009).

Despite numerous advancements in the reconstruction of Thanetian environments, the chronology of sea-level fluctuations remains debated. Two alternative eustatic curves, one proposed by Haq and Al-Qahtani (2005), who updated a previous version of Haq et al. (1987), and another one proposed by Miller et al. (2005) and later updated by

* Corresponding author. P.O. Box 7333, Rostov-na-Donu, 344056, Russian Federation.
E-mail addresses: ruban-d@mail.ru, ruban-d@rambler.ru (D.A. Ruban), svzorina@yandex.ru (S.O. Zorina), clintc@hawaii.edu (C.P. Conrad).

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Kominz et al. (2008), do not coincide well. Because of the potential for local tectonics or dynamic topography to affect stratigraphic measurements at any single continental location (Conrad et al., 2004; Moucha et al., 2008; Müller et al., 2008; Spasojevic et al., 2008, 2009; Conrad and Husson, 2009), the recognition and further verification of global eustatic changes requires an interregional correlation of regional records (Ross and Ross, 1985; Embry, 1997; Jacquin and de Graciansky, 1998; Hallam, 2001; Efendiyeva et al., 2008; Zorina et al., 2008; Kominz et al., 2008; Ruban et al., 2009). Previous studies proved the efficacy of this tool for the Paleocene (Zorina and Ruban, 2008a,b). In particular, a provisional comparison of data from the eastern Russian Platform, the Arabias platform, and Southern Australia permitted a recognition of common events despite their diachroneity. The present paper attempts to apply an interregional comparison of documented Thanetian shoreline shifts for a number of regions with well-established chronostratigraphic frameworks. Our main target is to examine the Thanetian global-scale transgressive–regressive (T–R) cycles.

2. Materials and methods

This study is based on an analysis of seven exceptional regional records. Only those regions with well-established chronostratigraphic frameworks and interpreted sedimentary successions are considered. They are the eastern Russian Platform, Northwestern Europe, Northwestern Africa, Northeastern Africa, the Arabian Platform, the northern Gulf of Mexico, and Southern Australia (Fig. 1). Representative sections of Thanetian deposits are documented in each of them (some examples are given on Fig. 2). Previously interpreted and already published information was preferred for the purposes of this study (Table 1). Meanwhile, some regional T–R cycles were deduced directly from the available stratigraphical charts (see Table 1 and Supplementary Information for sources of information and explanation of interpretations). One interesting observation made by our study concerns the quality of information in the available sources. When many T–R interpretations are published, some may be relatively poor in resolution, may mix different patterns of basin dynamics, or may not contain sufficient data. This potentially complicates the correlation of data across global domains, but remains important for discerning eustasy, which governs globally-coherent T–R cycles, from regional uplift and subsidence.

In the late Paleocene, the eastern Russian Platform experienced epeirogenic deformations, including swell inversion, arch growth, etc. (Nikishin et al., 1999). In the reference area of the Uljanovsk-Saratov Basin, the last inversion event occurred at the Cretaceous/Paleogene boundary, i.e., well before the Thanetian, when this area remained rather "stable". Somewhat similar geodynamics prevailed in Northwestern Europe, where local vertical tectonic motions superposed an otherwise "stable" regime (e.g., Vandycke et al., 1991; Vandenberghe et al., 1998). In particular, opening of the North Atlantic and reorganization of the rift system in the North Sea affected this region (Ziegler, 1975; Hardenbol et al., 1998). Both Northwestern and Northeastern Africa remained stable in the late Paleocene, but were affected by uplift in northern Africa and rifting in central Africa (Guiraud et al., 2005; Swezey, 2009). The Arabian Platform experienced subsidence coupled tectonic deformations at the plate margins (Sharland et al., 2001). Subsidence continued in the northern Gulf of Mexico during the late Paleocene, although the Laramide orogeny on the continent was an important control on terrigenous influx and, thus, sedimentation (Salvador, 1991; Zarra, 2007). Sediment loading itself favored an activation of fault tectonics and salt mobilization (Salvador, 1991). Southern Australia was evidently affected by the spreading to the south of the continent (McGowran et al., 1997), and new modeling indicates subsidence and an influence of subduction along the nearby West Pacific margin (DiCaprio et al., 2009).

The dispersed positioning (Fig. 1) and diverse tectonic settings (see above) of the study regions should prevent correlated changes in dynamic topography or local tectonics from affecting all records. For example, downwelling related to the Neo-Tethyan subduction zone could affect only a few the chosen regions or their parts, including the Arabian Platform, Northeastern Africa, and the northern portion of Northwestern Africa (Fig. 1). Similarly, different lithologies (Fig. 2) should prevent major biases linked to preferential preservation of shoreline shifts in particular kinds of sedimentary rocks. Thus, coherent T–R cycles among all, or even most, records should constrain a globally consistent pattern. To minimize the influence of local tectonic activity on transgressions and regressions, regions with relatively "stable" tectonic settings were preferred for the purpose of this study; tectonically active regions even with sufficient data and well-justified chronostratigraphy are therefore omitted. The New Jersey margin, where shoreline shifts provide exceptional constraints on sea-level changes (Van Sickel et al., 2004; Miller et al., 2005; Kominz et al., 2008), is also omitted in this comparison, but for another reason. Transgressions and regressions are shoreline shifts by definition (Catuneanu, 2006). Although they may reflect changes in the basin depth (i.e., what are called sea-level changes *sensu stricto*), this is not a rule (Catuneanu,

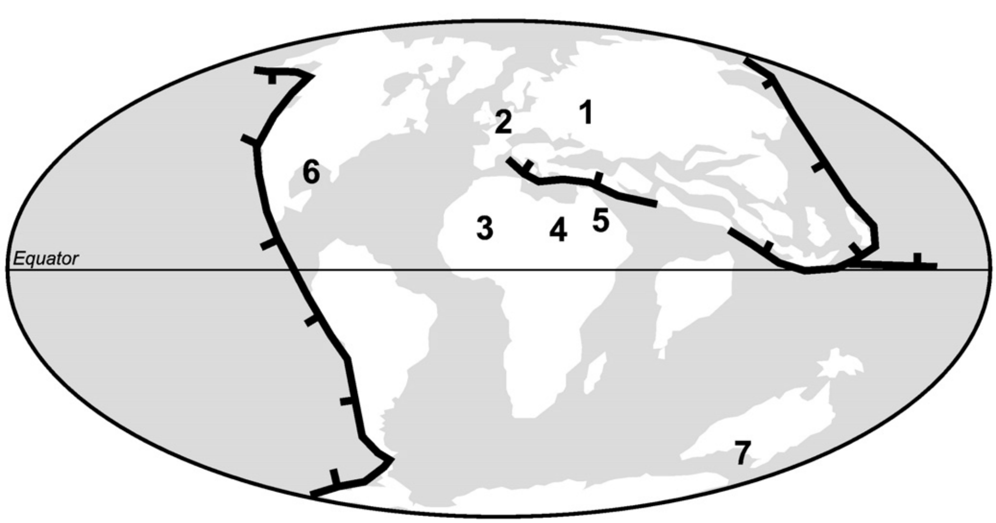

Fig. 1. Location of studied regions in the Paleocene. Plate tectonic map is simplified from Scotese (2004). Main active subduction zones are shown schematically (see www.scotese.com for reference). Regions: 1 — eastern Russian Platform, 2 — NW Europe, 3 — NW Africa, 4 — NE Africa, 5 — Arabian Platform, 6 — northern Gulf of Mexico, 7 — Southern Australia.

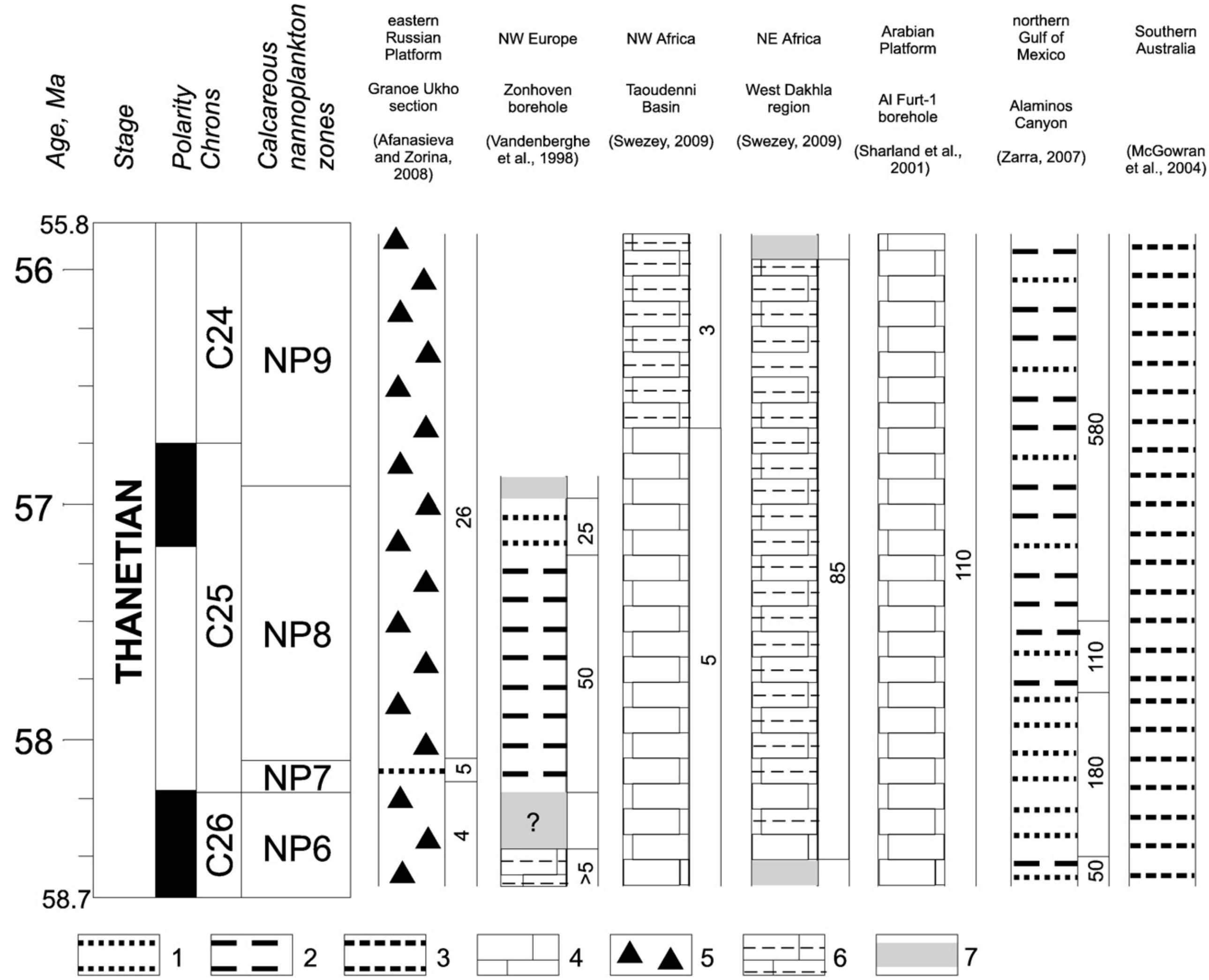

Fig. 2. Selected reference Thanetian sections of studied regions (very general correlations are possible). Approximate thickness (in meters) is indicated along each column. Lihologies: 1 — sandstones and all coarse siliciclastic, 2 — shales (and mudstones for the northern Gulf of Mexico), 3 — undifferentiated siliciclastics, 4 — carbonates (limestones), 5 — diatomites, 6 — marls and carbonate mudstones, 7 — hiatuses. More lithologic data are available within the sources of information (see Table 1). Chronostratigraphy after Gradstein et al. (2004) and Ogg et al. (2008). Ages of lithologic units represented at the Zonhoven borehole are corrected according to notes of Vandenberghe et al. (1998) given in their Fig. 8 (p. 130). Lithologic information for Southern Australia is given schematically on the basis of descriptions in McGowran et al. (2004). This figure provides only general information about the regional sedimentary records (individual lithologic columns cannot be used to recognize transgressions and regressions by definition). Larger data sets were used for T–R cycle interpretations (see Table 1 and on-line Supplementary Information).

2006; Ruban, 2007). Thus, sea-level changes measured accurately on the New Jersey margin (Van Sickel et al., 2004) are not suitable for a direct interregional comparison of T–R cycles. For the same reason, we do not focus on changes in water depth, which might or might not be related to the observed transgressions or regressions.

We have performed the interregional correlation as accurately as possible on the basis of the modern chronostratigraphy of the Thanetian Stage, which was summarized by Gradstein et al. (2004) and then updated by Ogg et al. (2008). We account for possible uncertainties in this chronostratigraphic framework (e.g., Westerhold et al., 2008), as well as possible errors in regional datings, within our interpretations. Our interregional correlation of regional records permits the recognition of common transgression and regression events and, thus, the identification of global-scale T–R cycles. This paper emphasizes transgressive and regressive trends and cycles, but does not treat their magnitudes, which can only be constrained by making detailed shoreline reconstructions at finely spaced time intervals. This is not yet possible for all regions. However, we do focus on the T–R cycles of the same (usually 2nd) order, which are, therefore, approximately comparable in duration and strength. Moreover, a synthesis of our data permits us to judge which transgressive/regressive patterns are of relatively larger or smaller magnitude in their overall appearance.

3. Results

3.1. Common patterns of the T–R cycles

An attempted correlation of regional T–R cycles gives a rather unexpected result (Fig. 3). No common transgressions or regressions can be recognized among the seven regions, but instead only decorrelated "noise" is observed. The eastern Russian Platform and the northern Gulf of Mexico are characterized by regressions at the beginning and the end of the Thanetian, whereas a transgression peak occurred in these regions somewhere in the mid-Thanetian with a certain degree of diachroneity. In other words, these regions were embraced by only one complete T–R cycle during the stage. Regressions, which peaked in the late Thanetian in Northwestern Europe, Northeastern Africa, and Southern Australia, were preceded by relatively long-term seaward shoreline migrations. In Northwestern Africa, a transgression lasted throughout the entire stage, whereas a similarly long-term regression is known from the Arabian

Table 1
Sources of information used in this study for our interregional correlation of Thanetian T–R cycles (detailed data on each region can be found in the sources indicated below). Cycles of the same magnitude (commonly 2nd-order) from all regions are only accounted to make the data consistent and, thus, sufficient for a comparison.

Region	Main source(s)	Type of information from the main source(s) used in this study	Stratigraphic control on T–R interpretation	Resolution of T–R interpretation	Additional source(s)
Eastern Russian Platform	Zorina and Afanasieva (2006), Afanasieva and Zorina (2008), Zorina and Ruban (2008b)	Interpreted T–R cycles (on the basis of facies analysis and maximum flooding surfaces identification)*	Diatom- and nannoplankton-based biozonation, polarity chrons; regional data are presented along the numerical time scale	Moderate	Distanov et al. (1970), Sahagian and Jones (1993), Akhmetiev and Beniamovskij (2003), Afanasieva (2004), Oreshkina and Aleksandrova (2007), Zorina and Ruban (2008a)
Northwestern Europe	Hardenbol et al. (1998), Michelsen et al. (1998), Vandenberghe et al. (1998)	Interpreted T–R cycles, coastal onlap record, sequence stratigraphic surfaces (on the basis of detailed sequence stratigraphic constraints)*	Foraminifera- and nannoplankton-based biozonation, polarity chrons; regional data are presented along the numerical time scale	Moderate	Van Eetvelde and Dupuis (2005), Ogg et al. (2008)
Northwestern Africa	Guiraud et al. (2005)	Unconformities, lateral facies distribution (interpretations of T–R cycles are based on observed changes in the lateral distribution of marine and continental facies and on the spatial extent of hiatuses)**	Only general chronostratigraphic scales are available (Guiraud et al. (2005) presented sedimentary records of various NW African basins on charts, where the Thanetian stage can be traced)	Low	Swezey (2009)
Northeastern Africa	Guiraud et al. (2005)	Unconformities, lateral facies distribution (interpretations of T–R cycles are based on observed changes in the lateral distribution of marine and continental facies and on the spatial extent of hiatuses)**	only general chronostratigraphic scales (within the numerical time scale) are available (Guiraud et al. (2005) presented sedimentary records of various NE African basins on charts, where the Thanetian stage can be traced)	Low	Swezey (2009)
Arabian Platform	Sharland et al. (2001), Simmons et al. (2007)	Lateral facies distribution, sequence stratigraphic surfaces (interpretations of T–R cycles are based on the documented sequence boundaries, on maximum flooding surfaces, and on the observed spatial extent of non-deposition)**	Foraminifer-based biozones, various chronostratigraphic data (referred, but not specified by Sharland et al., 2001 and Simmons et al., 2007); regional data are presented along the numerical time scale	High	Le Nindre et al. (2003)
Northern Gulf of Mexico	Zarra (2007)***	Coastal onlap record, 2nd-order sequence stratigraphic surfaces (interpretations of T–R cycles are based on the documented sequence boundaries, on maximum flooding surfaces, and on observed lateral distribution of facies)**	Foraminifera-based biozons, polarity chrons; regional data are presented along the numerical time scale	High	Galloway et al. (1991), Salvador (1991), Xue and Galloway (1993), Xue and Galloway (1995), Galloway et al. (2000)
Southern Australia	McGowran et al. (1997, 2004), McGowran (2005)	Interpreted 2nd-order T–R cycles (on the basis of facies sequence stratigraphic constraints, observed lateral distribution of marine facies and the spatial extent of hiatuses)*	Foraminifer-based biozonation, polarity chrons; regional data are presented along the numerical time scale	Moderate	DiCaprio et al. (2009)

* For these regions, previously attempted and published (in the main source(s)) interpretations of T–R cycles are utilized for the purposes of this study.
** For these regions, interpretations of T–R cycles are made in this study on the basis of information from the sources (see Supplementary Information for details).
*** Only high-resolution coastal onlap record (with no accounting of data on deep-water deposition) was used from this source.

Platform; neither transgressive nor regressive peaks are recorded in either of these two regions.

Thus, no common T–R cycles can be recognized in the Thanetian, at least in its early and middle parts. We should note that regressive peaks or trends are documented in 6 of 7 study regions during the late Thanetian (Fig. 3), with Northwestern Africa providing the lone exception. This may indicate a global-scale event of relatively large magnitude. However, the onset time of these regressions differed greatly. In the eastern Russian Platform and the northern Gulf of Mexico, the transgression peak was reached in the mid-Thanetian, whereas this occurred in the Selandian or around the Selandian/Thanetian boundary in Northwestern Europe, Northeastern Africa, the Arabian Platform, and Southern Australia. It is evident that regressions are slightly better correlated than transgressions, which does not match Hallam's (2001) observations for the Jurassic.

3.2. Comparison with eustatic curves

In the absence of dynamic topography or local tectonics, T–R cycles considered either individually or together should reflect eustatic changes, as global sea-level rise leads to the landward migration of the shoreline, and vice versa. Thus, it is useful to compare our seven distributed T–R trends to changes in global sea level during the Thanetian as documented by Haq and Al-Qahtani (2005) and Kominz et al. (2008). Both records are relatively poor in time resolution (however, this does not exclude a potential for comparison, because both curves, if even their authors aimed at long-term changes, indicate fluctuations during the Thanetian), and both do not coincide (Fig. 3). Haq and Al-Qahtani (2005) show a rise and then a highstand during the early-middle Thanetian and fluctuations in sea level during the late Thanetian, with a slight trend towards a rise. Kominz et al. (2008) noted three falls in the early-middle Thanetian superposed on a generally regressive trend, and a significant rise at the end of the stage. In other words, the curve by Haq and Al-Qahtani (2005) indicates a trend of eustatic rise throughout the entire stage, whereas that of Kominz et al. (2008) indicates a trend of eustatic fall interrupted by a rapid rise at the very end of the stage. Analysis of these trends is especially important for the purposes of the present study in order to reconcile the frequency of the documented T–R cycles and associated global fluctuations, i.e., to avoid problems with differing time resolution.

Both eustatic curves indicate little change in global sea level during the early-middle Thanetian, which may be consistent with our observation of an absence of global-scale T–R cycles. However, the lack of common trends between eustatic curves and the measured T–R cycles is enigmatic. Furthermore, the common regressive episode that we detected for the late Thanetian contrasts to the Haq and Al-Qahtani (2005) curve, but may correspond to the culmination of the long-term regressive trend constrained by Kominz et al. (2008). Thus, neither eustatic curve is fully supported by data from these 7 tectonically-“stable” regions. It is sensible to remember critical remarks expressed by Miall (1992) and Hallam (2001) with regard to the curve by Haq et al. (1987) as well as those by Müller et al. (2008) regarding the importance of dynamic topography for the Miller et al. (2005) record.

4. Discussion

4.1. Climatic control on T–R patterns

A total absence of globally correlated T–R cycles in the early-middle Thanetian requires an explanation. The first possibility concerns palaeoclimate: the Paleocene is regarded as a “hothouse” interval without major glaciations; temperatures increased by ~2 °C during the Thanetian (Zachos et al., 2001). An absence of continental ice precludes a glacial control of the global eustatic changes, which Clark et al. (2002) demonstrated can occur (e.g., at 14.7 ka for meltwater pulse 1A) at rates up to 40–50 mm/yr (equivalent to 40–50×10^3 m/myr, which is unsustainable for more than about 1 kyr). Furthermore, although thermal expansion of seawater can induce transgressions that occur as rapidly as ~50 mm/yr (e.g., Miller et al., 2005; Archer, 2008), the overall magnitude of rise from thermal expansion is only about 2 m/°C (Harrison, 1990), or about 4 m for the 2 °C of Thanetian warming. Thus, the absence of common transgressive trends thoughout the Thanetian is consistent with a relatively stable and warm paleoclimate during this stage that did not exert a significant globally-coherent influence on sea level.

4.2. Global tectonic controls on T–R patterns

Global-scale tectonic processes affect eustatic sea level by changing the “container” volume of the ocean basins, but typically induce only gradual sea-level change. For example, Rowley (2002) postulated nearly steady rates of lithosphere production and destruction during the last 180 Ma, which argues for sea-level stability. Other studies have suggested significant time-dependence of lithosphere production rates (e.g., Conrad and Lithgow-Bertelloni, 2007), and even an enduring slowdown during the Cenozoic and Cretaceous (Becker et al., 2009; Seton et al., 2009) that would lead to a decrease in ridge volume and thus a drop in sea level. However, both Xu et al. (2006) and Müller et al. (2008) reconstructed a gradual ageing of the oceanic floor lasting throughout the Paleogene, and estimated average rates of eustatic sea-level drop of only 1 m/myr (Müller et al., 2008) to 2–4 m/myr (Xu et al., 2006), although individual time periods exhibit near stability of sea level. Given the significant uncertainty in age reconstructions (Xu et al., 2006), it is not currently possible to resolve the tectonic contribution to sea level during a time period as short as Thanetian. However, the general tectonic trend of gradually decreasing sea level punctuated by periods of stability is consistent with the results of our interregional comparison of T–R cycles (Fig. 3).

Global tectonic processes can also induce eustatic change by increasing or decreasing the area of the ocean basins or by changing the average seafloor depth via sedimentation, emplacement of volcanic bathymetry, or changes in the component of topography that is dynamically supported by mantle flow. Again, rates of eustatic change are expected to be slow. Sedimentation and changes in average continental area and elevation together may produce only about 2 m/myr of sea-level change (Whitehead and Clift, 2009). Furthermore, the most significant ocean-area reducing event was the India–Asia collision (Harrison, 1990), which initiated at ~50 Ma (Zhu

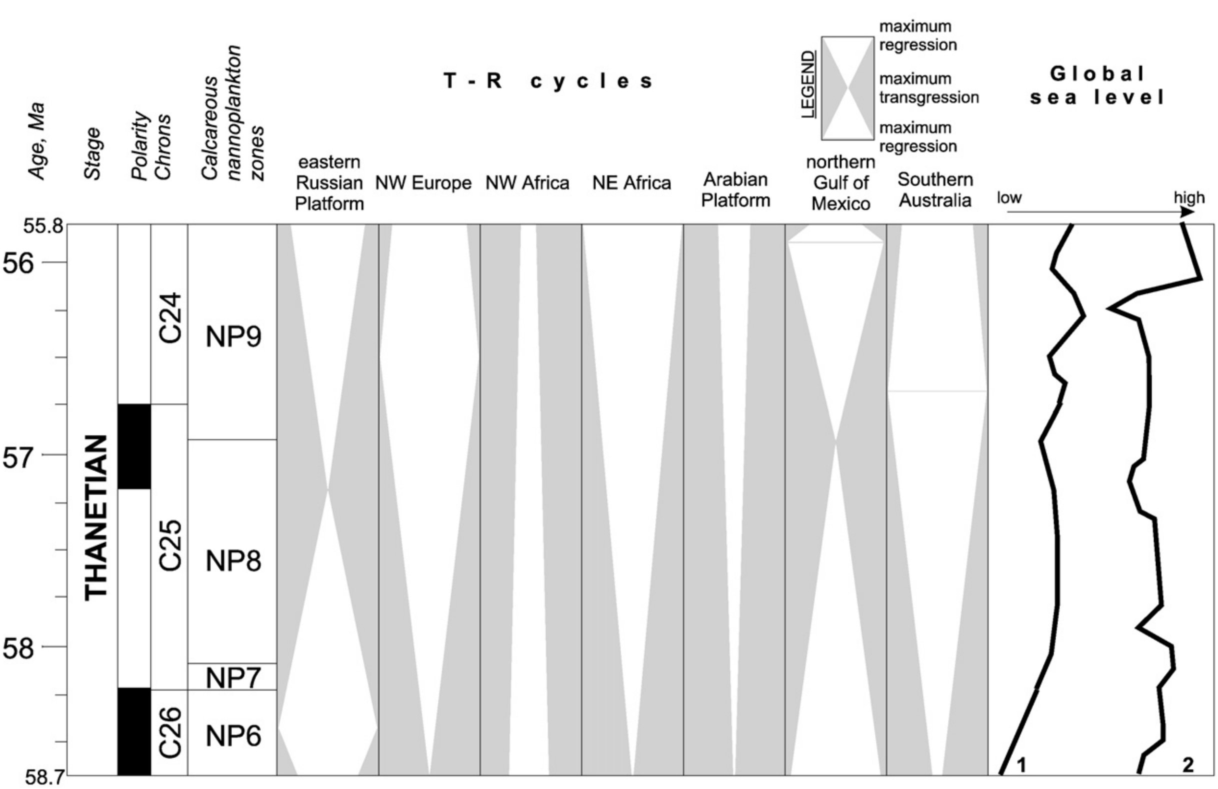

Fig. 3. Correlation of the Thanetian regional T–R cycles and global sea-level curves (1 — after Haq and Al-Qahtani (2005), 2 — after Kominz et al. (2008)). Note the relatively low resolution of the global sea-level curves. Table 1 lists sources of information. Chronostratigraphy after Gradstein et al. (2004) and Ogg et al. (2008). Westerhold et al. (2008) present a slightly different time scale.

et al., 2005), and thus after the Thanetian. Hallam (2001) and Zorina et al. (2008) suggested that mantle plume activity could regionally uplift continents and their margins, and, thus, induce regressions. The late Thanetian regression, which is documented in the present study, can be linked to the emplacement of the North Atlantic Large Igneous Province (LIP), which occurred contemporaneously (Saunders et al., 1997; Wignall, 2001; Jolley et al., 2002). However, this regression is traced even in such remote regions like the Arabian Platform and Southern Australia (Fig. 3), where the influence of the noted mantle plume activity is improbable. Alternatively, the emplacement of LIPs in the oceanic domain should decrease basin volumes and thus trigger eustatic rises of up to 5 m within a few myr. A tally of LIP volumes on the current seafloor suggest only 100 m of eustatic rise during the past 140 Ma, and no significant emplacement events during Thanetian time (Müller et al., 2008). Finally, dynamic topography can change the average bathymetry of the seafloor at rates that maximally produce only 1 m/myr of sea-level change (Conrad and Husson, 2009).

Taken together, global tectonic processes can be expected to produce at most about 7 m/myr of eustatic change during the Thanetian (namely, 4 m/myr from ridge volume changes, 2 m/myr from sedimentation, and 1 m/myr from dynamic topography), and only if all possible mechanisms are maximally producing change with the same sense. Although an upper bound, this relatively slow rate of eustatic change is consistent with our observation of only small coherent eustatic change during the Thanetian. Furthermore, all of the above-described global tectonic processes can be expected to operate on timescales much longer than the Thanetian, and therefore would not be useful for explaining the short-time scale T–R cycles observed at any individual margin (Fig. 3).

4.3. Dynamic topography as a control on regional T–R patterns

Vertical motion of individual shorelines caused by local tectonics or dynamic topography (the latter is especially important when tectonically "stable" regions are considered), may significantly influence the local expression of T–R cycles, regardless of any eustatic changes occurring at the same time. On a continental or basin-wise scale, several authors have demonstrated how long-wavelength dynamic topography induced by time-dependent mantle flow can uplift or depress continents relative to the sea surface (Lithgow-Bertelloni and Gurnis, 1997; Lithgow-Bertelloni and Silver, 1998; Conrad and Gurnis, 2003; Heine et al., 2008; Spasojevic et al., 2009). For eastern North America (Müller et al., 2008; Spasojevic et al., 2008) and Southern Australia (DiCaprio et al., 2009) this dynamic topography has been shown to govern continental-scale transgression and regression events because it induces local rates of uplift or subsidence that are faster than rates of global eustatic change.

By examining the time-dependence of global mantle flow, Conrad and Husson (2009) showed that mantle flow associated with lower mantle density heterogeneity can produce changes in dynamic topography that generate uplift or subsidence rates of 5–10 m/myr, correlated over thousands of km. Density heterogeneity in the upper mantle, however, can produce faster changes in dynamic topography because it is closer to the surface and resides is a lower-viscosity region of the mantle that can change more quickly. Conrad and Husson (2009) estimated that upper mantle density heterogeneity can induce uplift rates as high as 50 m/myr. Even faster uplift rates (100 to 250 m/myr) have been recently documented for sub-equatorial Africa and have been attributed to dynamic topography from upper mantle upwelling (Al-Hajri et al., 2009). If these rates of local uplift are faster than rates for globally occurring eustatic change, then a globally distributed set of individual T–R records will be dominated by regional dynamic topography, rather than global eustatic change.

Glacial melting and seawater thermal expansion can induce rapid eustatic change of several mm/yr, which is capable of overwhelming the 5–50 m/myr maximum rates of vertical motion caused by dynamic topography. However, the climatic stability and relatively warm paleoclimate during the Thanetian suggest that significant climate-induced eustatic change is unlikely. We have discussed above how processes associated with global tectonics together can produce at most only about 7 m/myr of eustatic change. In any one location, these rates of global eustatic change may be overwhelmed by regional changes in dynamically supported topography, which can be several times larger in magnitude and uncorrelated between regions. Therefore, the simplest explanation for the lack of correlation between our globally distributed set of Thanetian T–R records is that these records primarily reflect the influence of local dynamic uplift or subsidence, and not eustatic change. This implies that eustatic change during the Thanetian was significantly slower than the 5–50 m/myr rates of vertical ground motion that can be produced by dynamic topography.

The stratigraphic data used in this study documents T–R events with durations shorter than the 2.9 myr duration of the Thanetian. By contrast, long-wavelength dynamic topography produced by lower mantle density heterogeneity changes rather slowly. For example, Conrad and Husson (2009) showed that mantle flow driven by lower mantle structures causes ~300 m deflections of the surface that are changing at ~5 m/myr, which implies a ~60 myr timescale for dynamic topography induced by lower mantle flow. This timescale is consistent with time-dependent changes in dynamic topography associated with the Farallon slab in the lower mantle beneath North America (Müller et al., 2008; Spasojevic et al., 2008, 2009). By constrast, shorter-wavelength density heterogeneity in the upper mantle produces smaller amplitudes of dynamic topography that grow or shrink at 50 m/myr (Conrad and Husson, 2009) or faster (Al-Hajri et al., 2009), which suggests timescales shorter than 6 myr for dynamic topography associated with upper mantle density heterogeneity. This timescale is consistent with estimates of ~5 myr or shorter for the development of up to 600 m of uplift above a rising plume head (Griffiths and Campbell, 1991), and with the post-Pliocene development of ~500 m uplift over ~1000 m length scales in coastal sub-equatorial Africa (Al-Hajri et al., 2009). Thus, the relatively short timescales of the temporally uncorrelated T–R cycles during the Thanetian suggest that they may have been induced by dynamic topography associated with upper mantle density heterogeneity.

4.4. Relative stability of early-middle Thanetian eustasy: a synthesis

If global sea-level remained relatively stable during the early-middle Thanetian, it is likely that regional and local tectonic movements such as dynamic topography and basinwide changes in sedimentary budget were able to locally alter relative sea level and, thus, to create very specific and dissimilar regional records. In such a case, a difficulty in establishing a correspondence between T–R cycles during the early-middle Thanetian even for tectonically-"stable" regions is not unexpected. However, when regional events overwhelm the global trends, is it possible to tell about eustasy at all? For periods of relatively stable sea level, such as the early-middle Thanetian, we conclude that it is not possible to constrain eustatic curves or to delineate global-scale T–R cycles even on the basis of some well-known and seemingly tectonically "stable" regions. A broad and precise correlation of dozens of regions with various tectonic regimes is required in order to reveal any weak global sea-level fluctuations that may have occurred during this time (and questions will remain over whether these fluctuations reflect global signals). Until this broad spectrum of records is in place, we can only speculate simply about a period of relatively stable sea level. In this respect, the relative stability of the curve constrained by Kominz et al. (2008) is perhaps more appropriate for the early-middle Thanetian, despite the fact that it is determined from only one location, itself subject to local subsidence.

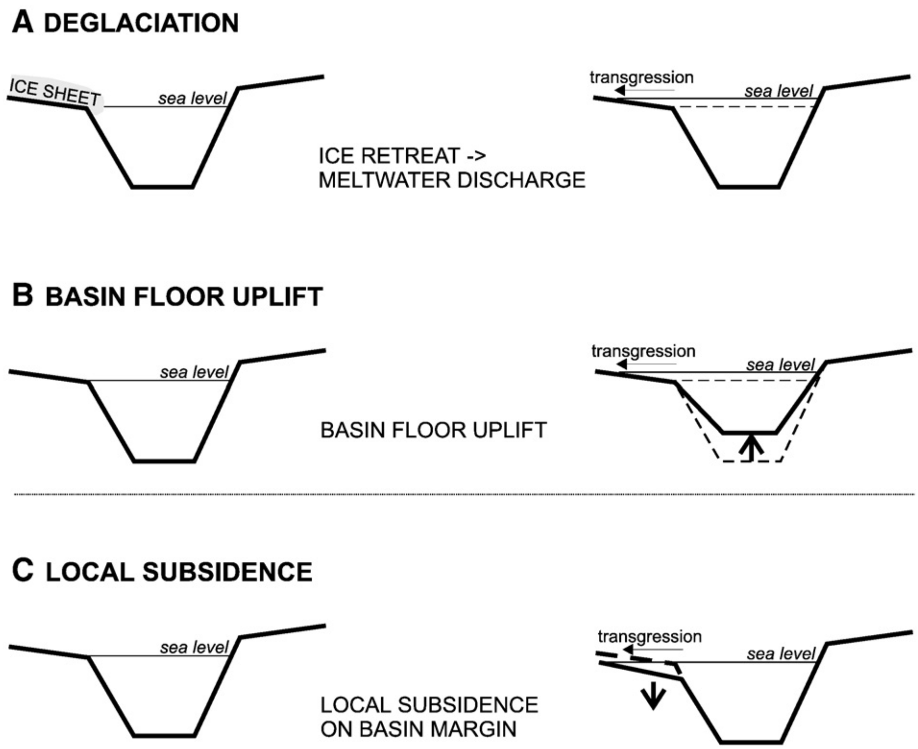

Fig. 4. Schematic representation of transgressions (which can be traced via landward shoreline shifts) driven by whole-basin processes that induce globally consistent transgression (e.g., water volume increase in A or container volume decrease in B), or by local processes (e.g., local subsidence in C) that induce only regional transgression.

4.5. Eustatic changes at the onset of the PETM

The PETM has been considered either a catastrophic culmination of gradual environmental changes that occurred since the early Paleocene, or an "unexpected" geological phenomenon. It is well known that the PETM produced a striking rise of the sea level (e.g., Miller et al., 2005; Kominz et al., 2008). Environments at the onset of this event remain debated. Although Zachos et al. (2001) show a warming trend through the Thanetian, Speijer and Wagner (Speijer and Wagner, 2002) and Speijer and Morsi (2002) suggested probable growth of cryosphere before the PETM and its further fluctuation. Their assumption is based, particularly, on observations of sea-level changes. Hallam and Wignall (1999) noted that sea level is important for any discussion of the PETM mechanism.

The results from our interregional correlation of the T–R cycles suggest a regressive episode in the late Thanetian. This episode followed an interval of above-mentioned eustatic stability, which lasted during the early-middle Thanetian. Undoubtedly, this late Thanetian regression is a reason to hypothesize a slight advance of glaciation in the high latitudes, which is consistent with the conclusion of Miller et al. (2005) and Galeotti et al. (2009) regarding the Cretaceous onset of Antarctic glaciations. Further detailed studies, including isotopic measurements, are required to confirm or reject cryosphere growth as the cause of late Thanetian regressions. If a minor late Thanetian glaciation episode is proved, cryospheric melting could contribute to PETM sea-level rise. In this case, one should conclude that the PETM was not a culmination of gradual global warming that occurred during the Paleocene.

5. Conclusions

An interregional correlation of the Thanetian T–R cycles allows us to make three important conclusions.

(1) No global-scale T–R cycles can be recognized in the early-middle Thanetian. This implies that both glacial (Fig. 4A) and container volume (Fig. 4B) mechanisms for global eustatic change were slow during this period of glacial absence and relatively constant tectonic processes. Instead, relative sea-level change at most continental margins was dominated by regional uplift or subsidence (Fig. 4C), likely caused by time-varying dynamic topography associated with upper mantle flow.
(2) A common regressive episode is established in the late Thanetian in the study regions. This episode might reflect ice growth.
(3) Neither of two recent eustatic curves (Haq and Al-Qahtani, 2005; Kominz et al., 2008) is fully verified, although the late Thanetian global sea-level fall suggested by Kominz et al. (2008) is confirmed by our correlation.

An absence of global-scale T–R cycles in the Thanetian brings up the question of whether regional sea level variations, such as those expected from dynamic uplift or subsidence of the Earth's surface by mantle flow (Moucha et al., 2008; Conrad and Husson, 2009), may commonly overwhelm global eustatic change in the geological past. Further investigations of other time intervals are required to fully answer this question.

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Appendix A. Supplementary data

Supplementary data associated with this article can be found in the online version, at doi:10.1016/j.palaeo.2010.05.040.

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