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Research paper

High-rate sea-level change during the Mesozoic: New approaches to an old problem

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Abstract

It is generally assumed that the effects of tectonism, sediment accumulation and compaction, and eustasy on accommodation change cannot be untangled on a regional scale. Hence, most reconstructions of past eustatic change have focused on global correlations. Here, this approach is questioned. In the time domain of few My and less, global cycle correlation is often unreliable and the effects of isostasy on sea level show strong provincial variations. In contrast, on a regional scale, cycle correlation is more precise and regional tectonism has predictable limitations. These considerations form the base for an assessment of the poorly understood causes of high-rate sea-level change in the Mesozoic. Three well-documented, regional, semi-quantitative sea-level curves are chosen here for reference (Jurassic of Britain, Cretaceous of Russia, Albian of Oman). The absolute numbers of amplitudes are open to question, but the order of magnitude (rates of several meters to tens of meters per My) is a robust property of all three curves even when considering large error bars. Variable mechanisms that cause these regressive–transgressive cycles are considered but rejected with exception of the controversial issue of glacio-eustasy or unknown mechanisms. Therefore, orbital forcing of glacio-eustasy is assessed. Freezing–melting cycles of higher-altitude ice shields can explain high-rate sea-level change with moderate amplitudes but conclusive evidence is buried beneath 3 km of East Antarctic ice. Thus, although several ambiguities of 'global' sea-level correlation are avoided by the regional approach, the causes of high-rate Mesozoic sea-level change remain poorly understood.

Keywords: Mesozoic; Sea level; Stratigraphic correlation; Bathymetry; Tectonism; Glacio-eustasy; Eutasty; Regional tectonism; Ice volume; Biostratigraphy

1. Introduction

In the presumed absence of major continental ice sheets, the mechanisms for high-amplitude, fast sea-level oscillations during the Mesozoic, perhaps the longest period of warmth during the Phanerozoic, are poorly understood (Frakes et al., 1992; Price, 1999). This as glacio-eustasy is the principal and vastly dominant driver in the domain of high-frequency, high-amplitude sea-level change. The publication of 'global' sea-level curves, including Mesozoic high-amplitude events (Vail et al., 1977; Haq et al., 1987), has fueled a considerable discussion on the validity of

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such standard curves (e.g. Rowley and Markwick, 1992; Hallam, 1998; Miall and Miall, 2001) and potential drivers of rapid sea-level change (e.g. Hsü and Winterer, 1980; Mörner, 1981; Lambeck, 1988; Cloetingh et al., 1985).

Commonly, the prerequisite of globally 'uniform' sea-level change is used to differentiate eustatic events from relative accommodation change due to regional and local tectonism (Cloetingh, 1986), and changes in sediment supply or environmental conditions (Schlager, 1993). Nevertheless, the elusive concept of globally uniform eustasy is questioned by geophysicists (Tamisiea et al., 2001; Potter and Lambeck, 2003), particularly when referring to glaciation–deglaciation cycles. Another problem lies in the field of global biostratigraphic time correlation. Depending on the time interval, ammonite zones and subzones, having the best long-distance correlation potential, are either provincial (e.g. Middle Jurassic: Branger, 2004), or global schemes (e.g. Hancock, 1991; Cenomanian: Hardenbol et al., 1998; Gale et al., 2002). During time intervals that allow for a global correlation of ammonite faunas, some apparently provide long-distance correlations with error bars that are as small as 50–100 ky (Gale et al., 2002). Nevertheless, when attempting to correlate Mesozoic 'eustatic' sea-level events on a global scale, with a time resolution of 1 My or less, and particularly during intervals of provincial ammonite faunas, the synchronous nature of many transgressive–regressive cycles remains notoriously difficult to prove (Aubry, 1991; Miall, 1992; Immenhauser and Scott, 1999).

The implication of these considerations is that the spatial dimensions for sea-level reconstruction and analysis must be chosen such that two conditions are satisfied: (1) the study area must be wider than the wavelength of local or regional tectonism, but (2) small enough for reliable time correlation of stratigraphic sequences.

Here, the Early Jurassic sea-level record of Britain (Hesselbo and Jenkyns, 1998), the Cretaceous sea-level record of the Russian platform (Sahagian et al., 1996), and the Albian curve of Oman (Immenhauser and Scott, 2002) are chosen for reference (Fig. 1). Two of these studies make use of recent shallow-marine seas to pinpoint the fair weather and storm wave base and translate these reference depths to well-documented sedimentologic, palaeontological and geochemical data from the sections or cores investigated. All three reference curves have in common that they suggest high-amplitude sea-level oscillations, correlated regionally across at least 500 km by means of stratigraphic markers, graphic correlation or well-established zonal–subzonal ammonite biostratigraphy (Hancock, 1991).

It is not attempted here to separate the various factors that, in their sum, result in relative sea-level change. Nevertheless, as the 500-km observational window (Fig. 1) is considerably wider than 50% of the greatest wavelength (anticline to anticline) of crustal lithospheric buckling (Cloetingh et al., 2002), tectonic causes for qualitatively uniform sea-level change across areas of this extent can be excluded. Hence, alternative mechanisms must be evaluated. From the different time frameworks, estimates of bathymetric change and the orders of magnitude of rates of sea-level change are deduced. These rates are placed against published rates of causes of sea-level change and the outcome is discussed.

A special focus is on modeled orbital forcing of Mesozoic glacio-eustasy, a particularly controversial aspect of climate research. From the three sea-level curves used (Jurassic of Britain, Cretaceous of Russia, Albian of Oman), only the Oman data are correlated to orbital modeling output (Immenhauser and Matthews, 2004). Therefore, the Albian example is used here as reference for the discussion of possibilities and pitfalls of this approach.

This manuscript has characteristics of a review article as it deals with longstanding problems including sea-level reconstruction and correlation (Miall and Miall, 2001), or the question of high-latitude/high-altitude ice shields during the Cretaceous greenhouse mode (Frakes et al., 1992). For this purpose, well-established regional curves are used and discussed in the context of published mechanisms that cause sea-level change (Dewey and Pitman, 1998). In contrast, the concepts of crustal lithospheric buckling (Cloetingh et al., 2002) or modeling of Cretaceous orbital forcing (Matthews and Frohlich, 2002), although now debated in the tectonics and the astro-chronostratigraphy communities, respectively, have perhaps not yet fully attracted the attention of those concerned about the reconstruction and interpretation of past sea-level change. The

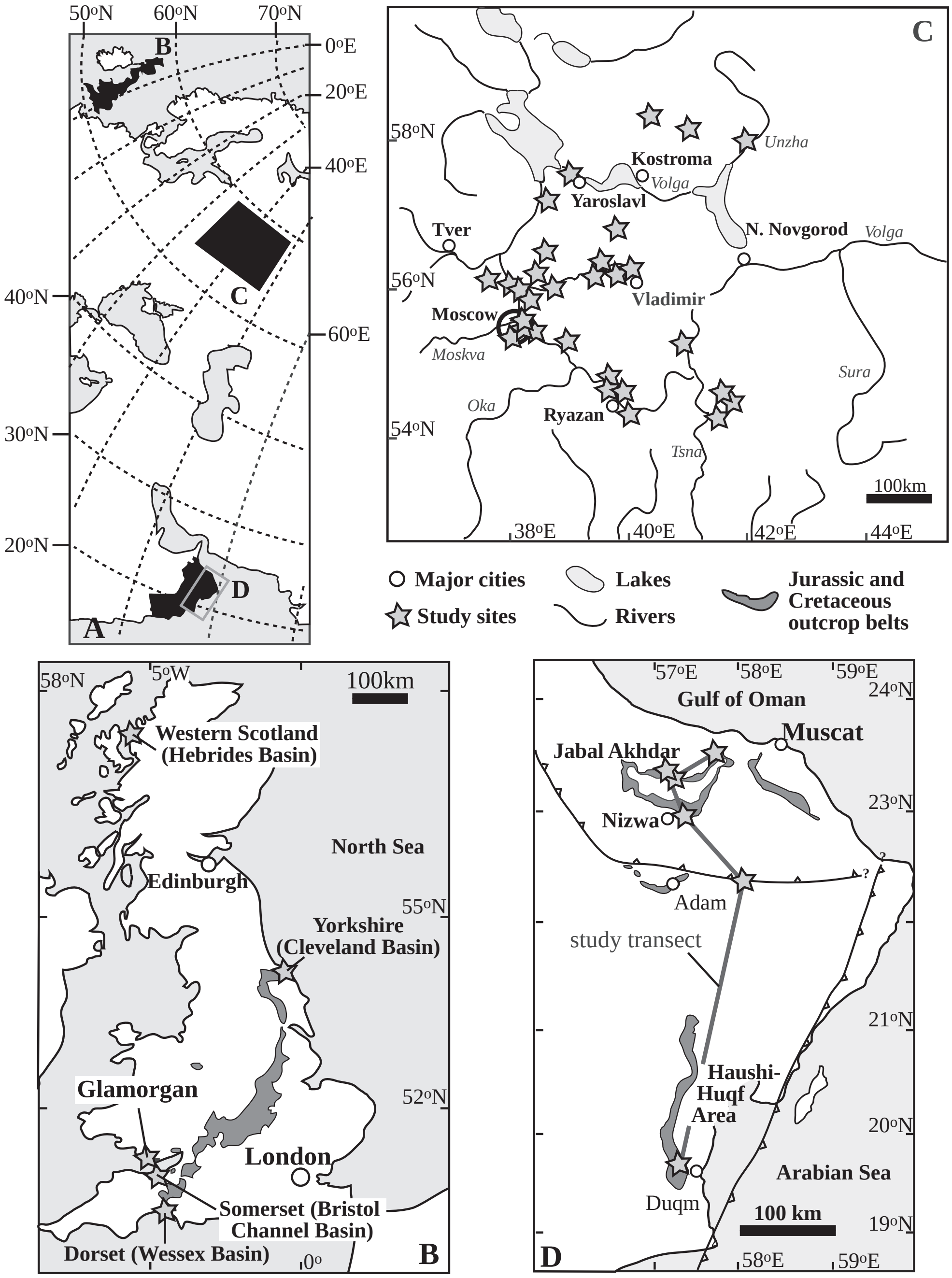

Fig. 1. (A) Geographic position of study areas. (B) Lower Jurassic sections of the British Isles (Hesselbo and Jenkyns, 1998). (C) Cretaceous sections and subsurface data from the Moscow depression (Sahagian et al., 1996). (D) Albian sections along the 500-km study transect in Oman (Immenhauser and Scott, 2002). Outcrop areas are shaded gray and localities studied indicated with a star.

pragmatic use of novel concepts from these fields of research might stimulate new discussion.

2. Case studies

2.1. Lower Jurassic of Britain

Hesselbo and Jenkyns (1998) re-investigated the classical sections of the marine Lower Jurassic of the British Isles to evaluate the synchronous nature of sea-level fluctuations recorded in different intrashelf basins (see Hallam, 1981, 1997 and references therein). For this purpose, all Lower Jurassic sections have been re-measured including from south to north, coastal exposures near Dorset, Somerset and Glamorgan, Yorkshire, and western Scotland (Fig. 1B). There, the predominantly siliciclastic and typically fine-grained Lower Jurassic sedimentary rocks were deposited in epicontinental basins with broadly similar tectonic histories and patterns of sediment accumulation (Fig. 2A). The large-scale cycles in these sections (Fig. 3A) were interpreted in the context of relative sea-level change and their synchronous nature across British basins was discussed in Hesselbo and Jenkyns (1998). Major erosional uncon-

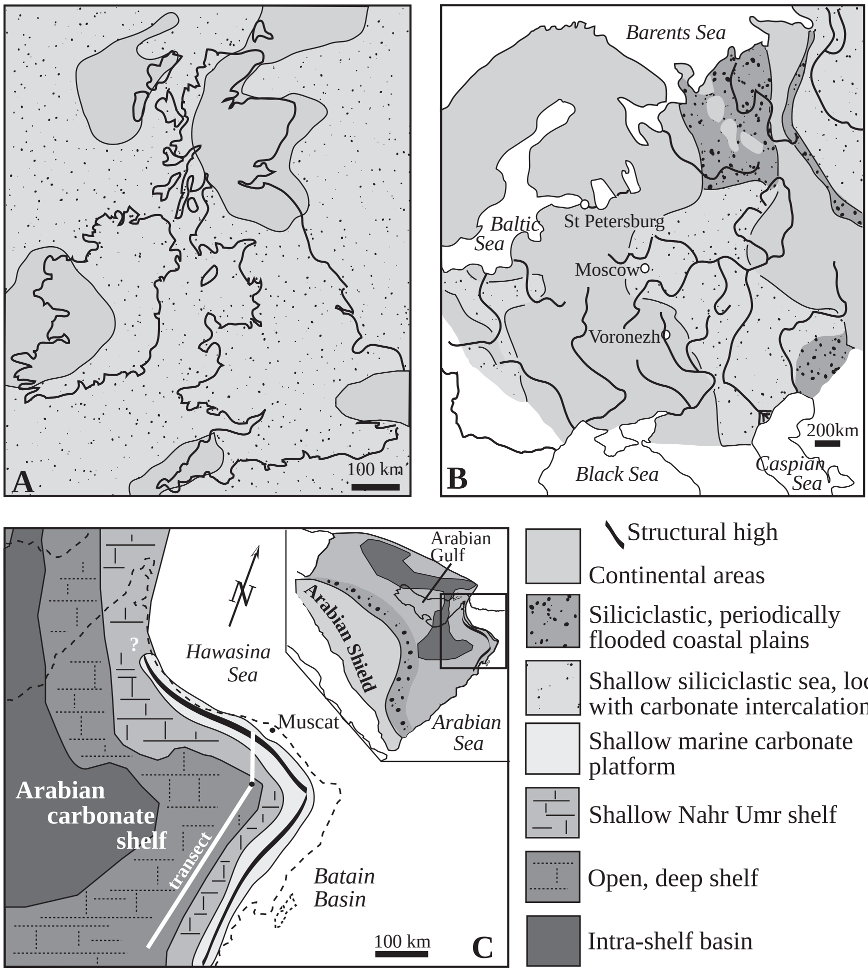

Fig. 2. (A) Palinspastic reconstruction of the Lower Jurassic of Britain (modified after Ziegler, 1990). (B) Reconstruction of Cretaceous facies belts in Russia (modified after Sahagian et al., 1996). (C) Albian palaeogeographic setting of Oman with indication of facies belts, main structural elements and study transect.

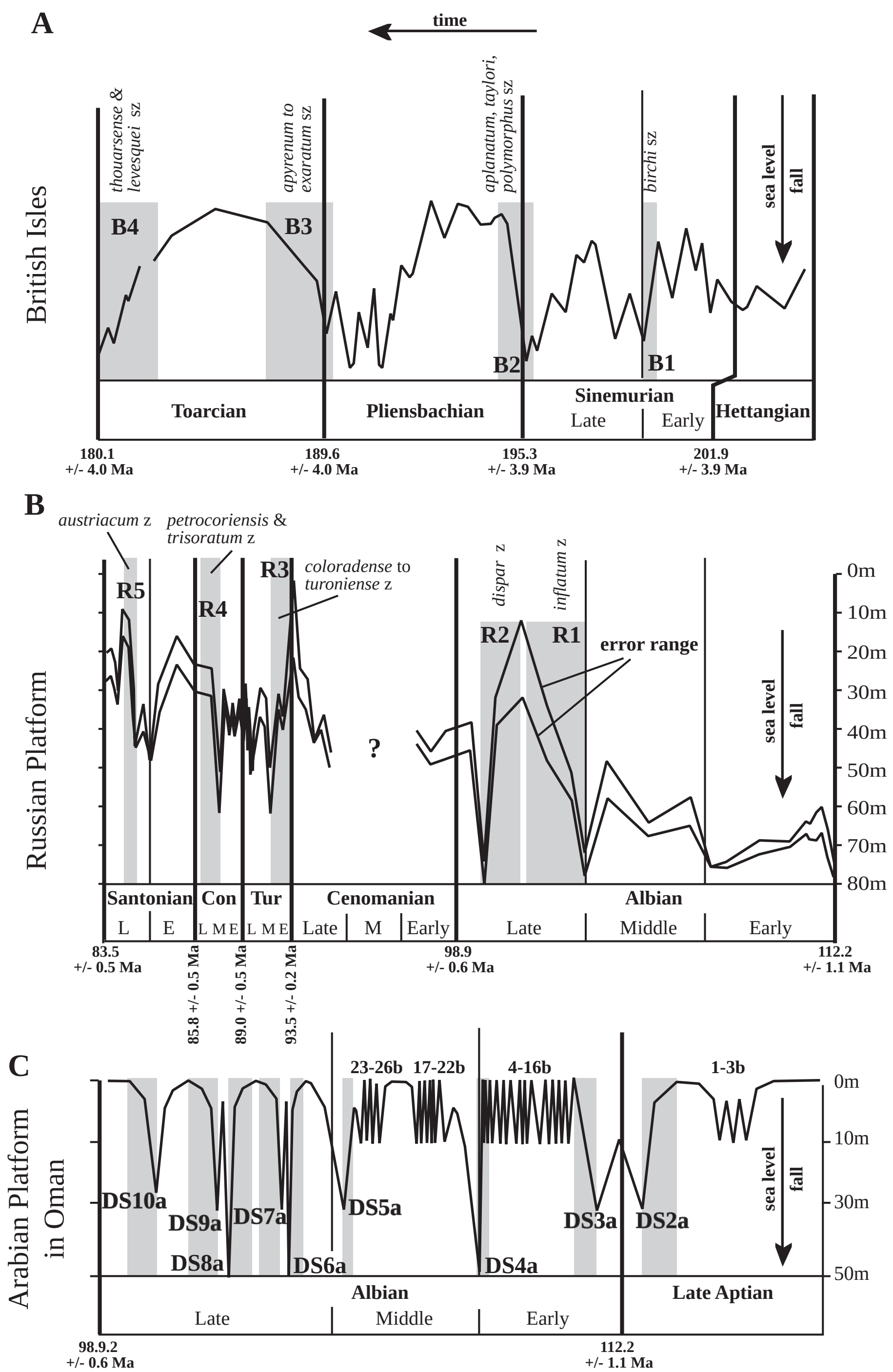

Fig. 3. Mesozoic sea-level reference curves. (A) Lower Jurassic sea-level reconstruction of the British Isles (after Hesselbo and Jenkyns, 1998). Sea-level events discussed in this contribution are shaded gray and labeled B1 through B4. Ammonite subzones are indicated. Ages from Weedon et al. (1999). (B) Albian through Santonian sea-level reconstruction from the Moscow depression (after Sahagian et al., 1996) with indication of ammonite zones. Sea-level events R1 through R5 are discussed. (C) Albian sea-level record of Oman (after Immenhauser and Scott, 2002). Sea-level cycles of intermediate and high amplitude DS2a through DS10a are discussed.

formities or facies successions indicative of minimal accommodation space in proximal areas were interpreted as sequence boundaries and are present in the mid-Sinemurian, the Upper Sinemurian, the mid-Pliensbachian, and the upper Toarcian. The resulting regional sea-level record is correlatable at the level of an ammonite subzone with that in other basins of continental northwest Europe (Hallam, 1997; Hesselbo and Jenkyns, 1998).

2.2. Cretaceous of the Russian platform (Moscow depression)

During the Jurassic and Cretaceous, the Russian platform was covered by a shallow epicontinental sea that extended from the Tethyan realm to the Arctic (Figs. 1B and 2B; Sahagian et al., 1996). The physiographic setting was that of a very low-gradient ramp that was repeatedly exposed and flooded. Marine deposition occurred with limited sediment supply from remote clastic sources alternating with subordinate carbonate beds. Sahagian et al. (1996) used the stratigraphic, palaeolontologic, sedimentologic and bathymetric record of the central part of the tectonically stable Russian platform and surrounding areas to construct a semi-quantitative sea-level curve for the Bajocian through the Santonian portions of which were used here (Fig. 3B). The sea-level record from different wells and outcrop areas was correlated using the well-documented ammonite biostratigraphy of the Russian platform. Relative sea-level curves were generated by backstripping stratigraphic data from well and outcrop sections distributed across an area of about 600×600 km (Fig. 2B). The reconstructed sea-level curve is characterized by a series of high-frequency sea-level events superimposed on longer-term trends. Major sequence boundaries are found in the Early and Late Albian, in the Turonian, Coniacian and the Late Santonian.

2.3. Albian of the Arabian platform (Oman)

From the mid-Permian through the earliest Turonian an extensive carbonate platform covered large parts of the Arabian craton (Figs. 1D and 2C; Sharland et al., 2001). During the Albian stage, several distinctive palaeo-depositional belts shaped this platform. Alluvial to coastal deposits rimmed the Arabian–Nubian Shield in the southwest, grading northeastwards into a shallow-marine carbonate shelf. In Oman, this carbonate shelf was superposed by a depression, here referred to as the 'Nahr Umr Basin'. Northwestwards, the Nahr Umr Basin probably opened into the larger and deeper Bab intrashelf basin of the United Arab Emirates (Fig. 2C; Immenhauser et al., 1999, 2001).

The Albian relative sea-level record in Oman was investigated at six localities along a 500-km transect extending from the northern slope of Jabal Akhdar to the southern Huqf area (Figs. 2C and 3C; Immenhauser and Scott, 2002). In a palinspastic reconstruction, this transect begins at the shallow-marine northern margin of Oman (Al Hassanat Formation), extends craton-wards into the Nahr Umr intrashelf basin (Nahr Umr Formation), and ends on the shallow shoulder of the Haushi–Huqf High at the eastern Oman margin. Within the studied sections, relative fall in sea level and the resulting emergence of the carbonate seafloor were recorded in subaerial exposure surfaces ('discontinuity surfaces' in Immenhauser et al., 1999, 2000). The subsequent relative rise in sea level is documented by normal marine carbonates overlying these surfaces. By combining the sea-level record from shallow-marine carbonate settings (Al Hassanat Formation) with the bathymetric record in the deeper setting of the Nahr Umr intrashelf basin (Nahr Umr Formation), the variable amplitudes of sea-level fluctuations are captured (Fig. 3C). The Oman curve shows major sequence boundaries in the Late Aptian and the Early, Middle and Late Albian. The reader is referred to Immenhauser and Scott (1999) for more details and a discussion of sea-level correlation between Oman and Russia.

3. Sea-level reconstruction

3.1. Amplitude of sea-level change: palaeo-water depth estimates

Various methods reconstructing palaeo-water depth have been proposed, but none is universally applicable (Hallam, 1967; Eicher, 1969; Benedict and Walker, 1978; Clifton, 1988; Brett et al., 1993). Most analyses use sedimentological, lithological, chemical, or palaeontological evidence and their typically association with specific water depths in modern marine environments, or palaeo-water depths are directly measured as distance below carbonate mound tops or platform breaks (Soreghan and Giles, 1999; Bahamonde et al., 2000).

Bathymetric changes in the Lower Jurassic of Britain are mainly interpreted from changing clay, or organic-rich mud-rocks, to sand fraction as the main siliciclastic constituents, but conglomeratic and oolitic limestones and Fe-rich sedimentary rocks are present at some levels too. As argued in Hesselbo and Jenkyns (1998), the sandstones have aspects typical of very shallow water settings. In contrast clays were deposited in greater water depts. Hallam (1997) evaluated the Early Jurassic palaeobathymetry changes using the Norton Sound in the Bering Sea (Johnson and Baldwin, 1986) and the German Bight (Aigner, 1985) as modern analogues. In the German Bight coastal sands pass into a mud–sand transition zone at about 7 m water depth, and this transition zone passes to shelfal mud at about 15 m. According to Walker and Plint (1992), the fair weather wave base in modern shallow, siliciclastic seas ranges from about 5 to 15 m and the storm wave base is about at 15 to 25 m depth. For the Lower Jurassic of Britain, this suggests depths of no more than about 20 m, where sedimentological evidence for storm activity is present, and a depth range of 50 to 100 m was proposed for fine-grained and finely laminated (sub-oxic) sections deposited beneath the storm wave base (Hallam, 1997). The different wave-base depths between different recent shallow seas exemplify the complex interaction of a variety of factors including the dimensions of the sea and the average and maximum wind-strength (fetch), the topography of the coastal area, properties of the sea-floor and sediment type.

For the sea-level reconstruction of the Russian platform, Sahagian et al. (1996) applied a semi-quantitative palaeobathymetry based on various types of sediments, sedimentary structures, mineralization, taphonomic conditions, ichnofacies, and bottom communities (Fig. 4). Furthermore, Sahagian et al. (1996) proposed that the present Baltic Sea forms an analogue to the Jurassic and Cretaceous Russian epicontinental seas. Observational and theoretical studies indicate that the modern Baltic Sea storm wave base is at 20–30 m (Zakharov, 1966). In general, the bathymetric parameters as used by Hallam (1997) for the Jurassic intrashelf basins of Britain and by Sahagian et al. (1996) for the Cretaceous Moscow depression of Russia are largely comparable because of the similar facies and palaeo-latitude.

In Oman, subaerial exposure surfaces overlain by normal marine carbonates were correlated along a platform to intrashelf basin transect and used as pinning points for the reconstruction of the Albian sea-level (Immenhauser and Scott, 2002; Fig. 5). The basic assumption is that marine biogenic carbonates do not build above sea level (Schlager, 1999). This implies that subaerial exposure surfaces in carbonates must be related to relative fall in sea level as opposed to decreasing accommodation space due to sediment aggradation. The relative bathymetry of each section is determined by differentiating (1) sediments deposited above the fair weather wave base (permanently agitated); (2) sediments deposited below the fair weather wave base, but above the depth of transient storm waves and storm-induced currents (intermittently agitated); and (3) sediments deposited below the storm wave base (no effective water agitation). Immenhauser and Scott (2002) used the present-day Persian Gulf and the Gulf of Carpentaria as recent analogues for the evaluation of palaeo-water depths. The present fair weather wave base in the Gulf of Carpentaria is at about 35 m, i.e. considerably deeper than the corresponding depths in northern European seas (5–15 m). Similarly, the effective storm wave bases in the Arabian Gulf are at about 40 m depth near islands and at about 70 m on the open ramp (see references in Immenhauser and Scott, 2002). The range of wave-base depths in modern shallow basins thus defines a range of uncertainty when attributing these parameters to Mesozoic sections.

3.2. Estimates of the rates of Mesozoic sea-level change

Each regressive–transgressive cycle can be described with two basic parameters: (1) The amplitude (magnitude) of relative sea-level fall or rise in meters, and (2) the duration of sea-level rise or fall in My, hence defining the rate (m/My) of sea-level change. Using the bathymetric parameters from recent shallow seas as analogues and the biostratigraphic framework from the three study areas, conservative estimates of the rates of selected Mesozoic sea-level events were made (events shaded gray in Fig. 3; Table 1). These specific cycles were chosen as they represent high-

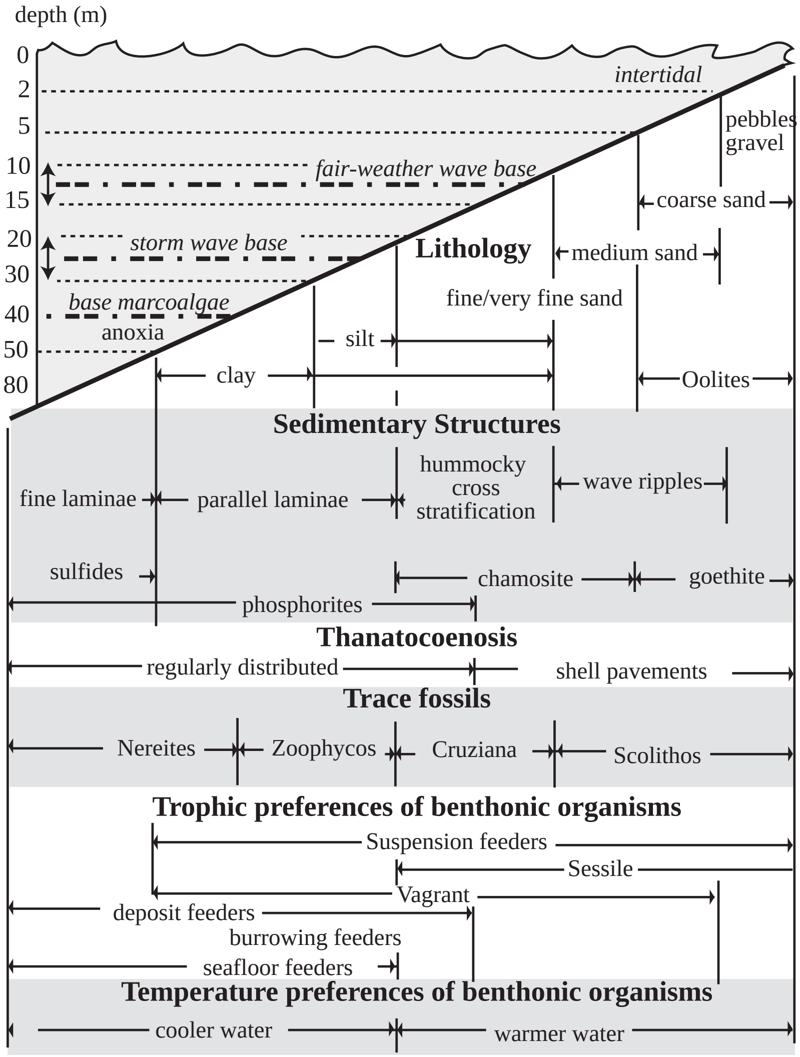

Fig. 4. Bathymetric scheme used for palaeobathymetry estimates of Cretaceous sea-level change in the Moscow depression (modified after Sahagian et al., 1996). The bathymetric parameters indicated in this scheme are largely applicable to the Jurassic sea-level curve of Britain. Refer to Sahagian et al. (1996) for more details.

amplitude, high-rate end-members of each curve. In the Lower Jurassic sea-level curve this is the Late Sinemurian (labeled B1), the Sinemurian–Pliensbachian boundary (B2), the late Pliensbachian (B3) and the end Toarcian (B4) sea-level rise/fall. In the Cretaceous sea-level curve of Russia, these are the Middle/Late Albian boundary event R1, the Late Albian event R2, the Turonian event R3, the Coniacian event R4 and the mid-Santonian event R5. In the Oman Aptian–Albian curve these are either rise or fall of the events DS2a through DS10a.

Here, the pragmatic approach of Hallam (1997) is applied in order to estimate rates of Lower Jurassic sea-level change in Britain (Hesselbo and Jenkyns, 1998) and the Cretaceous ones of the Russian platform (Sahagian et al., 1996). The assumption is made that within a given stage, ammonite zones (or subzones) are of equal duration. For example, the Sinemurian has a duration of 6.6 My (Gradstein et al., 1995) and 6 stages and 16 substages, respectively; thus the average duration of a Sinemurian stage is 1.1 My (0.4 My for a substage). A sea-level rise that is

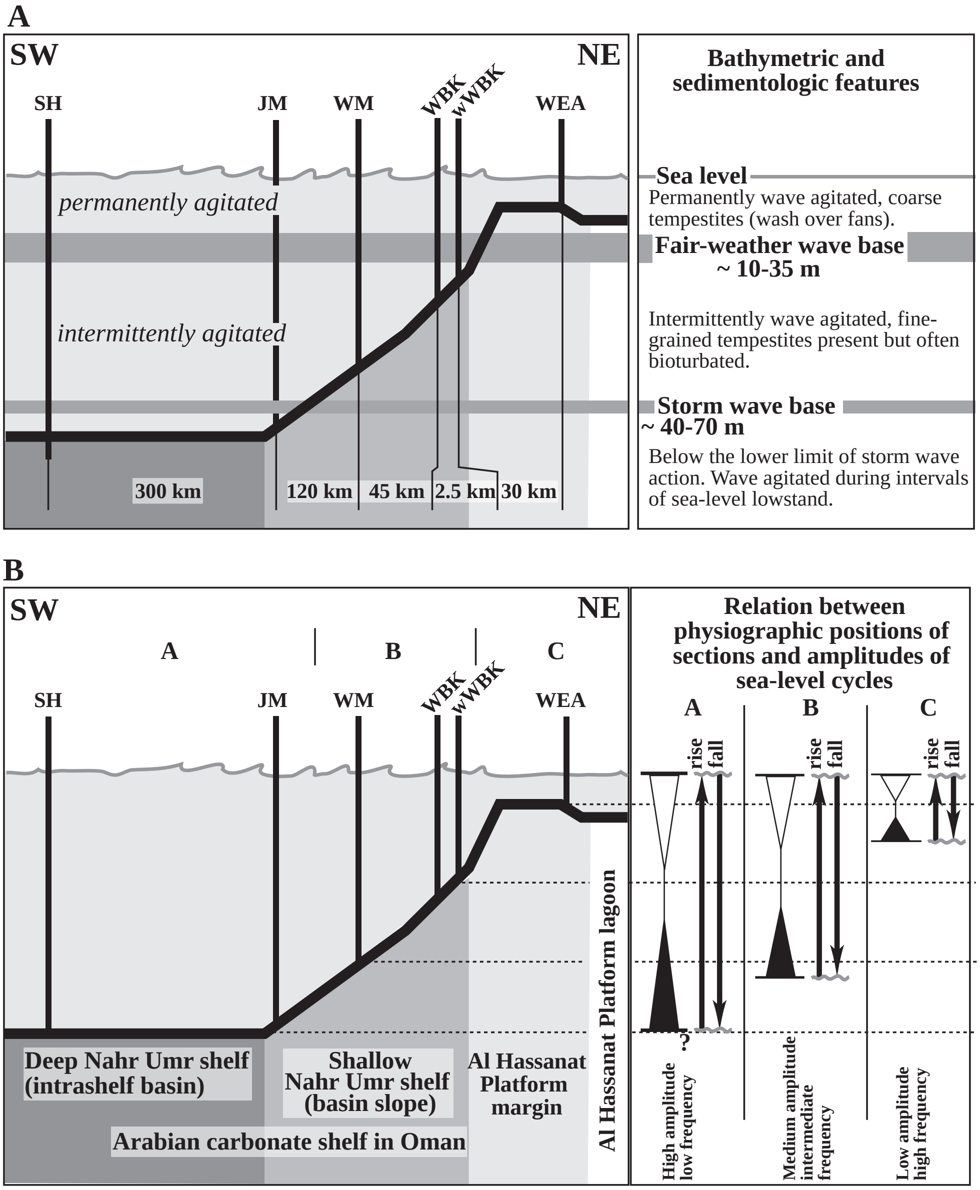

Fig. 5. Bathymetric scheme applied to the semi-quantitative sea-level curve of the Albian of Oman (modified after Immenhauser and Matthews, 2004). (A) Conceptual relation of bathymetry, wave agitation and bathymetry. The relative position of the sections and the lateral distance between sections is indicated. WEA=Wadi el Assyi; wWBK=west of Wadi Bani Kharus; WM=Wadi Mu'aydim; JM=Jabal Madar; SH=Southern Huqf. (B) Relation between average water depth and amplitude of high, medium and low-amplitude sea-level change. The relative position of sections is indicated.

recorded across 3 substages has therefore a duration of approximately 1.2 My (Table 1).

For the Albian of Oman, where a comparable ammonite biostratigraphy is lacking, another approach is chosen. The duration of the Albian is 13.3 My (Gradstein et al., 1995) and in Oman, nine regressive–transgressive events in the tens-of-meters scale are recorded. The average frequency of a regressive–transgressive cycle is thus 1.66 My but obviously this is a conservative estimate and some of these cycles

Table 1
List of selected sea-level events and related amplitudes, durations and rates

Sea-level events	Amplitudes (m)	Durations (My)	Rates (m/My)	Remarks and ammonite zones/subzones
Britain				
B1 fall	~10–30	~0.5	~22 to ~66	*birchi* subzone (average duration is 0.41 My)
B2 rise	~10–30	~1.2	~8 to ~25	*aplanatum*, *taylori* and *polymorphus* subzones (average duration/zone is 0.41 My (Sinemurian) to 0.38 My (Pliensbachian)).
B3 rise	~30 to 90	~3.5	~8 to ~25	*apyrenum* to *exaratum* subzones (average duration/zone is 0.38 My (Pliensbachian) to 0.63 My (Toarcian)). Hallam (1997) proposes either 30 or 90 m sea-level rise depending on the bathymetric parameters used.
B4 fall	~10–30	~3.2	~3 to ~9	*thouarsense* and *levesquei* zones
Russia				
R1 rise	~30	~1.5	~20	*inflatum* zone (~1.47 My)
R2 fall	~80	~1.5	~45	*dispar* zone (~1.47 My)
R3 fall	~30	~1.9	~15	*coloradense*, *nodosoides*, *turoniense* zones (average duration/zone is ~0.64 My)
R4 rise	~30	~1.4	~17	Portions of *petrocoriensis* zone and *trisoratum* zone (average duration/zone is ~0.875 My)
R5 rise	~35	~1.2	~30	Portions of the *austriacum* zone (average duration/zone is ~1.15 My)
Oman				
DS2a fall	~30	~0.8	~36	Average periodicity of Albian transgressive–regressive events is 1.66 My (i.e. from inflection point to inflection point)
DS3a rise	~30	~0.8	~36	
DS4a fall	~50	~0.8	~60	
DS5a fall	~30	~0.8	~36	
DS6a fall	~50	~0.8	~60	
DS7a rise	~30	~0.8	~36	
DS8a fall	~50	~0.8	~60	
DS9a rise	~30	~0.8	~36	
DS10a rise	~30	~0.8	~36	

were shorter whereas others were longer. Nevertheless, the average duration of either sea-level fall or rise is thus 50% of a cycle or about 0.8 My (Table 1).

Clearly, a considerable error bar must be attributed to the resulting rates of sea-level change but the absolute values as indicated in Table 1 are of little significance here. More important is the assumption that these rates represent geologically reasonable estimates in the correct order of magnitude, i.e. sea-level change in the domain of many meters to a few tens of meters per My. This clearly separates these events from such having rates of a few cm or dm per My.

3.3. Cycle correlation

Sahagian et al. (1996) and Hesselbo and Jenkyns (1998) make use of the ammonite zonal–subzonal system of Russia and Western Europe in order to date and correlate specific cycles. Ammonite zones and subzones, however, are either cosmopolitan (Gale et al., 2002) or provincial schemes (Branger, 2004) depending on the time window chosen (see discussion in Hancock, 1991). Sahagian et al. (1996) provides a correlation of European and Russian ammonite zones, but the ammonites of the Cretaceous intrashelf basins in the Middle East, for instance, are scarce and their correlation to the European system is difficult (Kennedy and Simmons, 1991). The correlation of expo-

Table 2
Causes of sea-level and related rates in comparison to estimated rates of Jurassic and Cretaceous sea-level change

Case studies and mechanisms of sea-level change	Rates (m/My)	Amplitudes (m)	Durations (My)	Dimensions (km)	Comments
Lower Jurassic of Britain (Hallam, 1997; Hesselbo and Jenkyns, 1998)	A few meters to several tens of meters	A few tens of meters	0.5 to 3.5	'Regional', i.e. more than 500 km; some events perhaps 'global'	Very detailed ammonite biostratigraphy. The bathymetric estimates are more difficult to establish. Recent analogues are used.
Cretaceous of Russia (Sahagian et al., 1996)	A few tens of meters	A few tens of meters	1.2 to 1.9	'Regional', i.e. more than 500 km; some events perhaps 'global'	Very detailed ammonite biostratigraphy. The bathymetric estimates are more difficult to establish. Recent analogues are used.
Albian of Oman (Immenhauser and Scott, 2002)	A few tens of meters	A few tens of meters	In average 0.8	'Regional', i.e. more than 500 km	Chronostratigraphy based on marker bed successions and graphic correlation. Subaerial exposure surfaces form solid bathymetric pinning points. Recent analogues are used.
Geoid deformation (glacio-hydro-isostasy)	~4 (not applicable to glacio-hydro-isostasy)	tens of meters	0.01 (near ice shields) to 10–20 globally	Regional to 'global'	Very fast at the margins of continental ice shields much slower due to long-term lithosphere–asthenosphere interaction.
Volume changes of mid-ocean ridge systems	~10 or less	Many tens of meters in extreme cases	tens	Regional to 'global'	
Hot spots, trap basalts, seamounts	~0.8	~30 m (Cretaceous)	tens	Regional to 'global'	
Regional to plate-scale tectonism	0.65–0.88	Many tens of meters in extreme cases	40–80	Regional to 'global'	
Plate margin deformation	1–10	>50		Regional	Plate margin deflection is a process that is expected to affect a 200–250 km wide zone of the plate margin.
Desiccation and flooding of areas below sea level	Not applicable	0.1 to perhaps 60		'global'	A possible area for Early Cretaceous desiccation and flooding of ocean basins is the South Atlantic.
Ocean sediment volumes	0.6 to 2.7	10–20		'global'	
Steric effects	~0.2	10 m or less		'global'	
Glacio-eustasy	Not applicable	Many tens of meters (up to 200 m)	0.01 to 0.1	'global'	

sure intervals between the Oman sections is thus based on graphic correlation and on stratigraphic markers (Scott, 1990; Immenhauser et al., 1999, 2000).

The problems related to biostratigraphic time resolution, the incompatibility of provincial biostratigraphic schemes for long-distance correlation, and the non-uniform behavior of different sea-level provinces (Potter and Lambeck, 2003) questions the validity of global correlation of sea-level cycles with durations of few My or less (Immenhauser and Scott, 1999). This is significant because the durations of sea-level change in the three case studies range from 0.5 to 3.5 My (Table 2).

4. Discussion of potential causes of Mesozoic sea-level change in the meters to tens of meters per My domain

In a general approximation, fluctuations in sea level are driven by either changes of the water volume in the world's oceans, or by changes in the shape of oceanic basins. The processes involved are discussed in, for instance, Lambeck (1988), Harrison (1990), Tooley (1993), or Dewey and Pitman (1998). Each process that affects sea-level has (within a certain range) characteristic amplitudes, periods and rates. These parameters are listed in Table 2.

Some of the factors listed in Table 2 (see also Dewey and Pitman, 1998) are not further discussed here. This because the resulting sea-level change has amplitudes or durations that are very different from those of the sea-level events under consideration (Table 2). This includes factors such as hot spots, trap basalts or seamounts, block scale (km to tens of km) basement movement, ocean sediment volumes or steric effects. Likewise, we do not consider plate-scale collisional (orogeny) events causing slow sea-level change with rates in the order of 0.65 to 0.88 m/My (Dewey and Pitman, 1998).

Flooding of desiccated Aptian South Atlantic, perhaps isolated from the worlds oceans at his time, could have caused ~60 m regressive events (Hsü and Winterer, 1980), but it is difficult to understand how these processes would cause rapid sea-level rise and this factor is not further considered here.

Geoid changes take place at a variety of time scales. The slow deformation of the Earth's long-wavelength components might perhaps lag behind present plate configuration by at least 100 My and involves processes that are too slow to be of relevance here (Davies, 1984; Gurnis, 1988). The effects of rapid regional geoid perturbation in the direct vicinity of continental ice shields (glacio-hydro-isostasy, e.g. Potter and Lambeck, 2003) are not considered either. This because otherwise the presence of low latitude, continental ice shields in Arabia (Albian) or intermediate latitude ice in Britain or Russia (Jurassic and Cretaceous) must be assumed. This is clearly not in agreement with modeled climate or field evidence.

Thus, the subsequent discussion focuses on three remaining principal mechanisms: (1) volume changes of mid-ocean ridge systems; (2) intraplate deformation; and (3) glacio-eustasy.

4.1. Volume changes of mid-ocean ridge systems and spreading rates

It has been argued that during the mid-Cretaceous, the ocean–crust production rates were exceptionally high (Larson and Kincaid, 1996) leading to a globally high sea level (Komintz, 1984). Dewey and Pitman (1998) calculated the effect on sea level of a change in spreading rate for the case of the present Pacific Ridge System. The resulting net rise in sea level by a 1 cm per 10 years increase in the spreading rate is 54 m.

Grötsch et al. (1993) for example, argued for a volcano-tectonic mechanism in the case of the high-amplitude, high-rate sea-level fall and subsequent rise in the Late Albian *Rotalipora appeninica*-zone. Similarly, Hallam (1997) interpreted the Pliensbachian–Toarcian boundary sea-level rise as recorded in Britain and elsewhere as perhaps driven by mid-ocean ridge volume increase, but no Early Jurassic oceanic lithosphere is preserved that could support or contradict this assumption. Moreover, as argued in Dewey and Pitman (1998), only the most extreme cases of spreading change cause sea level to rise with a rate of 10 m/My. The rate of sea-level change is most rapid directly after an increase in ridge spreading rate and then exponentially decays as a function of time, i.e. causing rapid sea-level rise and much slower fall.

In conclusion, extreme changes in spreading rates might cause sea-level rise in the order of 10m/My, but it seems difficult to assign rapid sea-level fall to this mechanism and evidence for unusual thermal activity at Jurassic and Triassic mid-ocean ridges is lacking.

4.2. Regional plate deformation

The deflection of thinned lithosphere at passive continental margins and at the margins of intrashelf basins is dominated by erosion, sediment loading, and thermal contraction (Fig. 6A). Based on these observations, Cloetingh et al. (1985) proposed a tectonic model for regional relative sea-level variations with rates of 1–10 m/My and magnitudes of many tens of meters. The model explains these sea-level changes, provided that horizontal stress fields

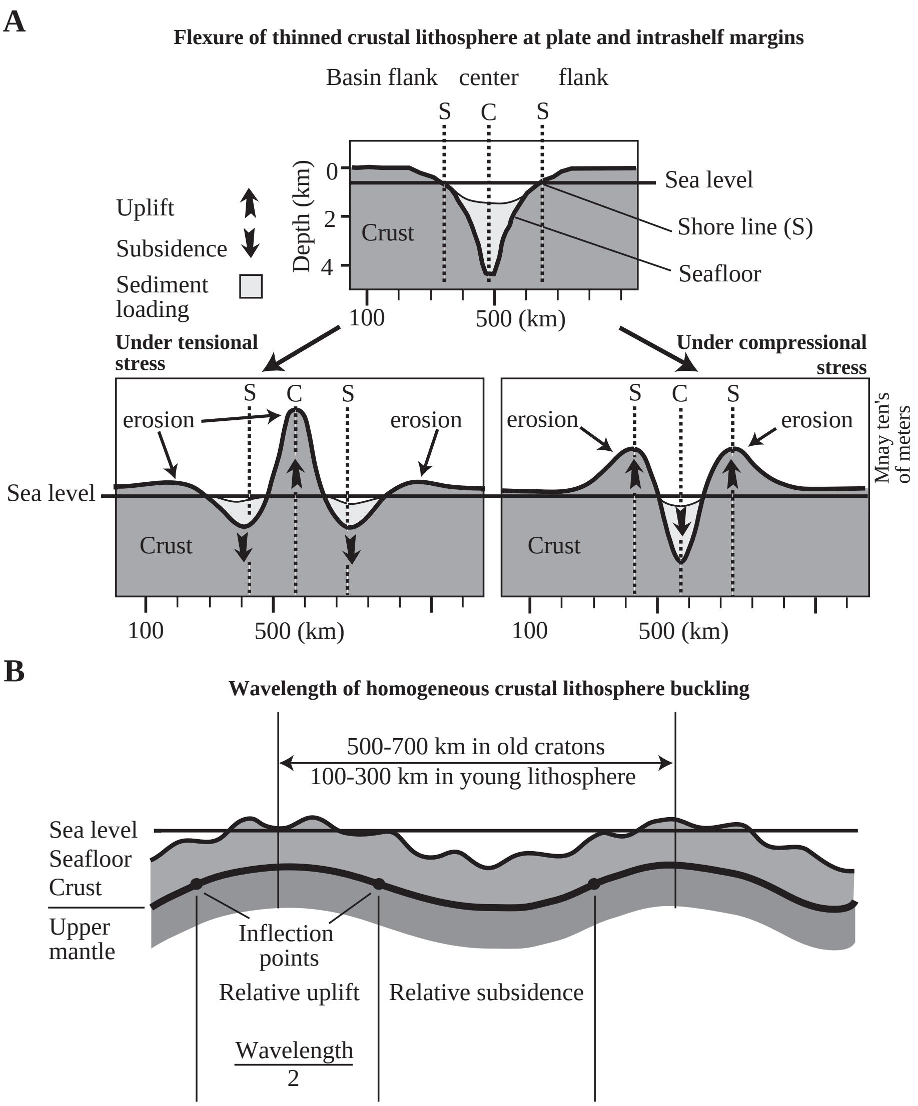

Fig. 6. Regional tectonism and relative sea-level change. (A) Flexure of thinned crustal lithosphere at intrashelf basin or plate margins. Basin flanks subside under tensional (left) or become uplifted under compressive stress (right). This is recorded as either unidirectional relative sea-level rise or fall (after Cloetingh, 1986). Initial positions of shore line (S) and basin center (C) are shown for reference. (B) Buckling of homogenous crustal lithosphere is expressed in different wavelengths and causes relative sea-level fall across half of the wavelength (anticlines) and relative rise in the synclines (after Cloetingh et al., 2002).

occur at continental margins, or at the margins of intrashelf basins, on geologic time scales. The concepts of Cloetingh (1986) have attracted considerable interest and specific sea-level events in the Haq et al. (1987) standard curve can be attributed to periods of plate tectonic reorganization. Similarly, Hallam (1997) argued that intraplate stress could explain the sharp sea-level fall and rise across the Triassic/Jurassic boundary in Britain and many other sections in Europe for instance.

Nevertheless, the deflection of plate margins and intrashelf basin margins has clear spatial and mechanical limitations that must be considered in this context. For instance, plate margin deformation is expected to affect a zone of perhaps 200 km but usually much less (Fig. 6A; Ravaut et al., 1993). Furthermore, lithosphere deflection is recorded as unidirectional relative sea-level fall at the basin margin (under compressive stress) and unidirectional relative sea-level rise in the basin itself, or the opposite under tensile stress (Fig. 6A). Thus, in order to explain several regressive–transgressive cycles at one locality, the assumption of repeated stages of compressive plate-boundary reorganization rapidly alternating with extensive stress must be made. This is mechanically possible but clearly not in agreement with geological evidence.

With reference to buckling (folding) of crustal lithosphere with uniform thickness (as opposed to the thinned crust at plate margins), Cloetingh et al. (2002) documented that the wavelength of lithospheric folding has spatial dimensions that must be considered when dealing with the effects of regional tectonism on relative sea level. In essence, the first-order wavelength of lithospheric folding is largest in ancient, stable cratons (500 to 700 km from anticline to anticline) but becomes much shorter in younger lithosphere (Fig. 6B). In a field study from Central Asia, Nikishin et al. (1993) document wavelengths of lithospheric folding in the order of 360 km superimposed by a smaller order wavelength of 30–50 km. The implication of this is that lithosphere buckling cannot explain sea-level change that is, within geologic error bars, correlated uniformly (in a qualitative manner) across 500 km or more. This is because across more than half of the maximum wavelength of lithospheric folding, flexural bending will cause relative sea-level rise at one locality and relative sea-level fall elsewhere (Fig. 6B). Furthermore, Dewey and Pitman (1998) argued that in stratigraphic sequences, in-plate stress is primarily recorded as local relative change in water depth, whereas this mechanism has a very limited impact on global sea level.

In summary, plate margin deflection and lithosphere buckling are reasonable causes for regionally limited unidirectional relative sea-level rise or fall. All other factors such as basement subsidence, block faulting, variable sediment accumulation or compaction rates will influence relative water depths at a given locality. Nevertheless, these mechanism are inherently non-systematic and local and cannot explain repeated regressive–transgressive events recorded qualitatively uniform across several hundreds of kilometers.

4.3. Water transfer between continental ice shields and ocean basins

The principal and vastly dominant mechanism in the domain of the high-frequency, high-amplitude, sea-level drivers is glacio-eustasy. Much has been debated about the possible presence or absence of continental ice shields in the Mesozoic and an adequate discussion of this topic is beyond the scope of this paper (see Francis and Frakes, 1993; or Price, 1999 for reviews).

Nevertheless, in strata of Late Cretaceous age from France and Canada, Plint (1991) and Malatre et al. (1998) recognize high-amplitude, high-frequency sea-level oscillations and infer a glacio-eustatic control. This would imply that a significant number of glaciations of sufficient size to affect global sea level occurred during the Mesozoic. Similarly, Stoll and Schrag (1996, 2000) or Weissert and Lini (1991) discuss geochemical evidence of glacial interludes in the Cretaceous. Moreover, Wiese and Voigt (2002) document faunal response in Europe pointing to glacial conditions or at least 'cold snaps' for the early Cenomanian, the late Turonian and the Coniacian.

These observations are contrasted by the fossil remains of middle and Late Cretaceous polar forests that thrived under a combination of light and temperature regimes that does not exist in our present world (Read and Francis, 1992). Similarly, studies on the Jurassic phytogeography of Siberian floras (Ziegler et al., 1993) indicate cold winters and temperate summers, implying polar conditions considerably warmer than today (see discussion in Price, 1999). Nevertheless, given that the Earth is a spherical body receiving solar insolation unequally over its surface and that heat transport in both atmosphere and hydrosphere has clear physical limits, it is difficult to explain how the polar zones could ever have been warm enough to melt all ice and snow there (Frakes and Francis, 1988). Quantitative palaeo-temperature estimates have lead Spicer and Parrish (1990) to infer that permanent ice was likely in Alaska above altitudes of 1200 m during the Cenomanian and above 1000 m during the Maastrichtian. Given the fact that the Transantarctic Mountains stretch across 3500 km, a relief difference of 1 km would allow for the existence of seasonally moderate climates in lower latitudes (i.e. intermontane basins and coastal plains with forests), contrasted by much colder average temperatures at higher elevations where substantial perennial ice sheets could exist.

In conclusion, the issue of Mesozoic glacial episodes and their impact on sea-level remains controversial. The field area most likely to prove or disprove the existence of Mesozoic ice sheets at higher altitudes in Antarctica, however, is presently buried beneath 3 km of East Antarctic ice. Nevertheless, the synergetic collaboration of climate modelers and geologists might perhaps lead to progress in this field of research.

5. Orbital forcing of climate and sea level—interaction between stratigraphers and modelers

In the last decades, two fundamentally different geosciences communities have approached the issue of palaeo-climatology and sea-level reconstruction. These are climate and tectonic modelers setting the processes and then viewing the results, and geologists reconstructing past climates and sea level from physical and chemical proxies attempting to extract the underlying processes. To test hypotheses, however, the modelers need boundary conditions and proxy values from geologists. This is where synergistic collaboration between modelers and geologists is most constructive (see discussion in Soreghan and Giles, 1999).

With respect to glacio-eustasy and climate change during the past few My, orbital forcing has been a successful hypothesis (e.g. Berger et al., 1999). Although several attempts are now under way to extend orbital time scales at least as far back as the Maastrichtian, many astronomers have little confidence in orbital forcing calculations on time scale beyond more than 35 to 50 Ma. For these longer time scales, problems concerning eccentricity and small variations in these frequencies can introduce large uncertainties (Berger and Loutre, 1994; Laskar, 1999). In order to asses these uncertainties, Matthews et al. (1997) evaluated the effects of specified displacements in the orientation of planetary orbits and proposed a trigonometric series expression from 65 to 190 Ma (Matthews and Frohlich, 2002). The trigonometric terms and their initial phase angles are derived by Fourier analysis (e.g. Berger et al., 1992; Berger and Loutre, 1994) from the numerical integration of the gravitational effect of each planet on all other planets, back through time (Laskar, 1990; Quinn, 1991).

In order to investigate the physical limitations of possible higher-altitude ice in high-latitude settings and orbital forcing of Mesozoic sea level, Immenhauser and Matthews (2004) applied the 'parametric forward modeling' (PFM) strategy of Matthews and Frohlich (2002), which inputs orbital forcing time series and outputs a calculated sea-level time series. The PFM applies this parameterization (see Matthews and Frohlich, 2002) to an ice-sheet area and solar insolation specific to that step in the calculation and thus calculates the size of the ice sheet for the next time step of the model. Differences in size of the ice sheet are taken as water to or from the global ocean, thus sea-level change.

By modeling an orbital forcing sea-level curve for the Albian, and combining it with stratigraphic modeling of sedimentation response to sea-level change, a comparison of the model output with the field-based Albian sea-level curve from Oman was possible (Fig. 7). The approach chosen is similar to the model–data comparison of for instance northern hemisphere ice volume over the last 3 My (e.g. Berger et al., 1999) although the error bars involved in modeling Mesozoic orbital forcing of the sea level are considerably larger. In addition, all of these model

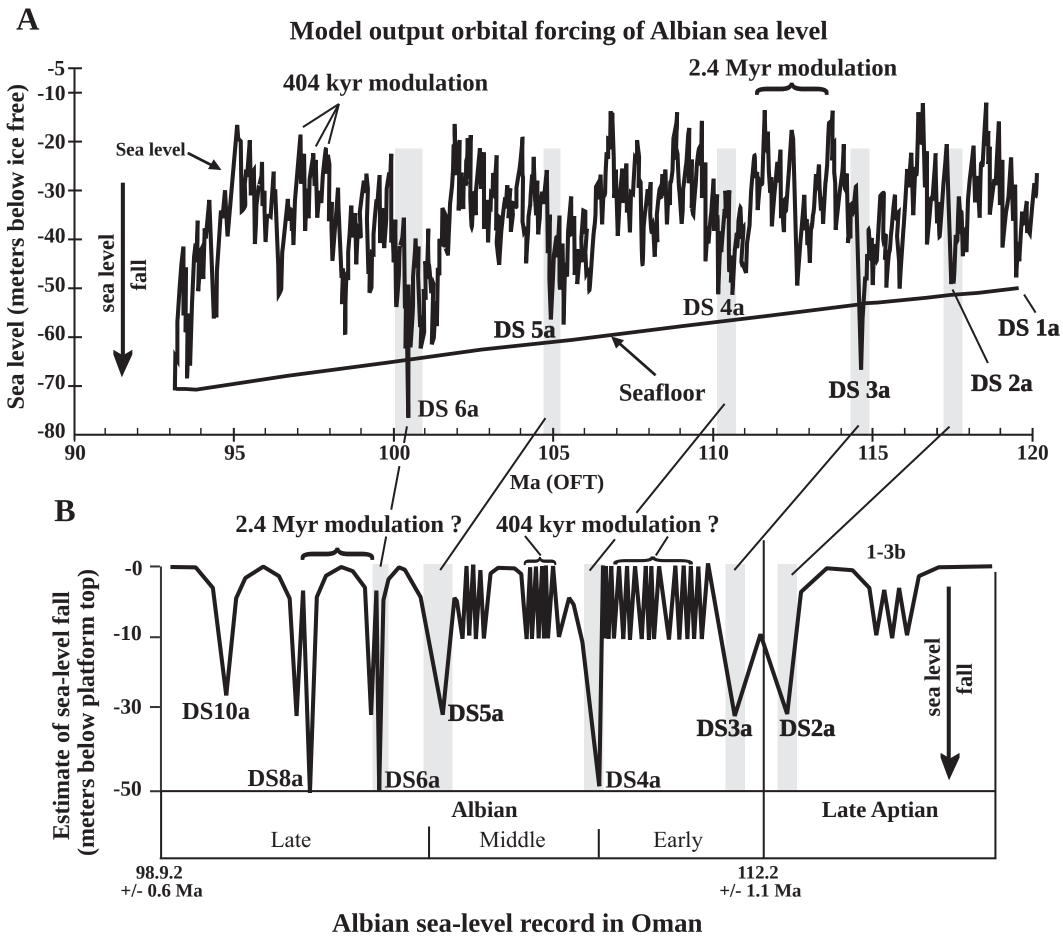

Fig. 7. (A) Model result depicting sea level and sediment surface as a function of time for the Wadi Bani Kharus section of the Albian of Oman (SimStrat PFM parameterization of Matthews and Frohlich, 2002; modified after Immenhauser and Matthews, 2004). The 404 ky and the 2.4 My climate modulations are indicated. Time scale is in Ma (OFT) according to SimStrat PFM parameterization that is not directly comparable with biostratigraphic ages (Ma). (B) The Albian sea-level record of Oman. Refer to Immenhauser and Scott (2002) for more details. DS=discontinuity surfaces pointing to sea-level fall and subaerial exposure of the carbonate seafloor.

approaches are limited because of model boundary conditions (Berger et al., 1998).

A principle observation is the identification of 404 ky and 2.4 My modulation (Fig. 7) as a major pattern of climate oscillations and thus sequence stratigraphy (Matthews and Frohlich, 2002; Immenhauser and Matthews, 2004). Biostratigraphic evidence from Oman might suggest that this pattern is present in the Albian sea-level record of the Arabian platform (Fig. 7; Immenhauser et al., 1999).

As shown in Fig. 7, the match of this data–model comparison of the Oman curve is expectedly not perfect but surprisingly good and this approach is definitely worth pursuing. With respect to model–data differences, the model suggests high-amplitude sea-level change in the case of discontinuity surfaces (DS) 3a and 6a, whereas field data suggest sea-level events 4a, 6a and 8a to have highest amplitudes. This might have several reasons, but one of the main data/model differences is that the model exclusively reflects orbital forcing, whereas the Albian rock record in Oman is expected to represent an array of climatic, tectonic and local sedimentological factors. See Immenhauser and Matthews (2004) for a detailed analysis of the model–data comparison.

Obviously, as it is the case for all climate models, the outcome of this data–model comparison adds no conclusive evidence to the controversial discussion on Mesozoic ice in general (Price, 1999). Nevertheless, the model approach allows for testing and quantifying some of the underlying parameters. This information, in turn, can be used to critically evaluate and improve working hypotheses or to reject them. For example, the amplitude estimates of the orbital forcing model output (80 m or less) slightly exceed the limitations of an Antarctic ice shield (Immenhauser and Matthews, 2004). This implies that, for instance, the many-tens-of-meters amplitudes of sea-level change proposed for the Cretaceous of Russia regional curve (Fig. 3B; Sahagian et al., 1996) represent either overestimates or that it must be concluded that other, unknown, mechanisms were active.

In summary, the model–data comparison suggests that the Oman sea level and the orbital forcing pattern triggering ice making–melting cycles have broad similarities. Given the unambiguous presence of fossil Mesozoic forests on Antarctica, allowing for the presence of higher-altitude ice only, the estimated amplitudes of Albian sea-level change in Oman require further explanation.

6. Summary and conclusions

Three well-constrained sea-level reconstructions (Lower Jurassic of Britain, Cretaceous of the Moscow depression, Albian of Oman) show tens-of-meters scale sea-level oscillations, correlated regionally across at least 500 km by means of stratigraphic markers, graphic correlation, or well-established zonal–subzonal ammonite biostratigraphy. Considering the limitations of global time correlation and the provincial effects of glacio- and hydro-isostasy, regionally uniform sea-level reconstructions are based on better evidence than attempted correlations of global eustasy in the time domain of 1 My or less. This under the prerequisite that regional tectonism as the main driver of relative sea-level change can be excluded.

Biostratigraphic data from the reference studies and bathymetric pinning points from recent shallow-marine seas place the rate of Mesozoic sea-level change in the meters to tens-of-meters per My domain. This is a robust property of these curves even when considering a wide range of uncertainty encompassed in palaeobathymetry reconstruction and time correlation. Of all known causes of sea-level change, only volume changes of mid-ocean ridge systems, lithospheric deformation (regional tectonism), and glacio-eustasy fall into this domain.

Thermal anomalies, causing rapid spreading at oceanic ridges, characteristically result in rapid sea-level rise and much slower fall, hence cannot explain high-rate regressive–transgressive cycles. Regional tectonism has clear mechanical and spatial limitations as deduced from tectonic models and field observations. Considering these limitations, an observational area of at least 500 km is more than the half of the wavelength of even the largest lithospheric folds. Thus, uniform and correlatable sea-level change across this distance cannot be explained by buckling of crustal lithosphere. Moreover, flexure of thinned lithosphere at basin margins causes unidirectional sea-level rise or fall at one locality and thus cannot explain regressive–transgressive cycles either.

With reference to Mesozoic glacio-eustasy, this topic remains particularly controversial due to contradicting evidence. However, the possibility of intermediate size, higher-altitude perennial ice shields at high-latitude settings should perhaps not be excluded. In order to assess the possibilities and limitations of ice melting–making, orbital forcing of glacio-eustasy is investigated as an explanation of the Oman Albian data. Numerous observational patterns of orbital forcing are recognized, but an exact match is not achieved. An implication of the model data is that the amplitude of the Albian sea-level change in Oman cannot be accommodated by freezing–melting cycles of spatially limited high-altitude ice shields on Antarctica. This implies that either the proposed amplitudes represent overestimates or that other unknown mechanisms are involved. Conclusive evidence, however, is presently buried beneath 3 km of East Antarctic ice.

Thus, although several ambiguities of ‘global’ sea-level correlation are avoided by the regional approach proposed here, the causes of high-rate Mesozoic sea-level change remain poorly understood. One might speculate that several mechanisms, causing sea-level rise or fall, coincided in time and thus aggravated each others effects; that the possibility of short glacial 'snaps' in the Mesozoic period of warmth is perhaps underexplored; or that unknown mechanisms were active that have no modern counterparts.

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