Albian sea-level cycles in Oman: the ‘Rosetta Stone’ approach

Adrian Immenhauser and Robley K. Matthews

ABSTRACT

The Albian Nahr Umr Formation of Oman (bounded by the Shu’aiba and Natih formations) recorded high-, intermediate, and low-amplitude regressive-transgressive events. In order to reconstruct these sea-level oscillations in a semi-quantitative manner, six measured sections from different bathymetric positions were investigated across a 500-km transect. Between these sections, discontinuity surfaces, indicative of sea-level fall and subsequent rise, were correlated. In order to assess the underlying causes of sea-level oscillations we compared the amplitudes and frequencies of the Oman events with published amplitudes and frequencies of factors that affect the water volume in the Earth’s oceans, or the shape of oceanic basins. We used the spatial array and field, chemical and petrographic properties of these rocks as a key for unraveling the unknown factors that caused sea-level change. We concluded that sea-level cycles in Oman either reflect unknown processes we do not understand, or changes in continental ice volumes. Thus, we give serious attention to the concept of orbital forcing and glacio-eustasy.

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 data-model comparison was possible. The match of our data-model comparison is expectedly not perfect but surprisingly good, and this approach is definitely worth pursuing. One of the main data-model differences is that the model exclusively reflects the orbital-forcing signal, whereas the Albian rock record in Oman is expected to represent an array of climatic, tectonic and local sedimentological factors. Nevertheless, the good data-model match supports the hypothesis that glacio-eustasy was the dominant driver of Albian sea-level change in Oman. This outcome might prove to be controversial in that it suggests a re-evaluation of the Cretaceous as a period of global, continuous climatic warmth, without significant ice shields at the poles.

This interaction has highlighted the many differences between the stratigrapher’s perspective and modeler’s perspective. There are good and bad points to both perspectives, which we point out and attempt to reconcile here.

INTRODUCTION

Natural variations in sea level are evident over a large range in time and space scales starting with diurnal tides and reaching to fluctuations in sea level with amplitudes of 100s of meters and frequencies of 10s of millions of years. Nevertheless, since the 19th century, sea-level research has revolved around three key issues:

1. What processes drive sea-level fluctuations?
2. How to separate the local, regional and global drivers of sea-level fluctuations?
3. How to reconstruct or predict, respectively, past or future sea-level change?

More recently, the relation of sea level and climate, particularly the prediction of future sea-level fluctuations and the impact of a potential sea-level rise on low-lying coastal regions has become a central focus of applied sea-level research (Intergovernmental Panel on Climate Change, 2001). The driving forces of present-day sea-level fluctuations, their relation to climate, and the complex superposition of different factors clearly are a matter of an ongoing turbulent debate (Lambeck, 1988; Pugh, 1993; Warrick, 1993; Intergovernmental Panel on Climate Change, 2001; Siddall et al., 2003).

The difficulties involved in sea-level reconstruction become even more evident when dealing with the fluctuations of ancient ocean surfaces that can only be indirectly deduced from the bathymetric...
interpretation of the rock record (Burton et al., 1987; Hallam, 1992). With reference to the elusive concept of eustasy (i.e. the global change of sea level throughout the world’s oceans), two fundamentally different views have emerged (Tooley, 1993). One view is based on large industry data sets including seismic stratigraphy, outcrop and well-log data, and from the 1970s onwards stressed the concept of a globally-uniform stratigraphic framework driven by eustatic sea-level fluctuations (Vail et al., 1977; Hallam, 1981; Haq et al., 1987; Hardenbol et al., 1998). This concept resulted in the publication of ‘global sea-level charts’ (Haq et al., 1987; Hardenbol et al., 1998).

The opposing view, mainly brought forward by geophysicists and tectonic modelers, questioned the existence of globally-uniform changes in sea level (e.g. Clark et al., 1978; Tamisiea et al., 2001). This concept has mainly resulted from the discovery of regional deformations of the geoid relief, and the variable distribution of the asthenosphere under oceans and continental margins following the loading or unloading of meltwater to ocean basins (Cloetingh, 1986; Mörner, 1987; Lambeck, 1988). The regionally variable expression of eustasy is empirically documented for the most recent glacial cycle and was modeled for time intervals of 10,000 years after the melting of a hypothetical ice shield (Clark et al., 1978). Within this time domain, different sea-level ‘provinces’ apparently react differently to loading and unloading of meltwater (Clark et al., 1978). This was subsequently confirmed by other workers (e.g. Tamisiea et al. 2001, and references therein) using tide gauge records, high-resolution numerical simulations and satellite-based mapping of ongoing sea-surface perturbations. Tamisiea et al. (2001) showed that ice mass variations on the Earth’s surface lead to highly non-uniform changes in relative sea level and of the Earth’s sea surface (geoid).

In the more distant past, i.e. the Neogene, Paleogene or Mesozoic, however, and when dealing with sea-level fluctuations with frequencies of one to several my, our ability to record such regional differences is limited. This is because of the considerable error bars related to the resolution of past time, the position of the ancient sea surface and its fluctuations with time. The problems involved become even more evident when focusing on (geologically) relatively short time slices in the order of about 1.0 my and less, i.e. when approaching the limits of biostratigraphic time resolution (Immenhauser and Scott, 1999).

The purpose of this paper is to evaluate the processes that triggered Albian regressive-transgressive events recorded in the Nahr Umr and Al Hassanat formations of Oman (Immenhauser and Scott, 2002). The sea-level cycles observed are present across a 500-km-long transect, i.e. they are of semi-regional significance (Figure 1). The potential local, regional and global drivers are presented and their applicability to the Oman sea-level curve is discussed. We consider the Oman Albian data set reviewed here to contain sufficient detail and geographic and environmental variability to make it a veritable ‘Rosetta Stone’ for unraveling processes that drive mid-Cretaceous sea-level change. We refer to this as the ‘Rosetta Stone Approach’ in analogy to the basalt plate inscribed by Egyptian hieroglyphs, Egyptian demotic, and Greek, which was found by a French officer of Napoleon’s engineering corps near the city of Rashid (Rosetta), Egypt, in 1799. The tablet, known as the ‘Rosetta Stone’, provided the translation between Egyptian hieroglyphs and the Greek language, and unlocked the meaning of the messages written by the ancient Egyptian civilization. In this paper, we evaluate the hypothesis that the interpreted amplitude and frequency of the Albian sea level of Oman constitute a ‘Rosetta Stone’ for the ‘geological hieroglyphs’ in the stratigraphic record.

After finding problems with numerous alternative processes, we give serious attention to the hypothesis of orbital forcing and glacio-eustasy (Matthews and Frohlich, 2002; Mattner and Al-Husseini, 2002), a particularly controversial aspect in the mid-Cretaceous ‘greenhouse world’.

REGIONAL GEOTECTONIC SETTING

From the mid-Permian through the earliest Turonian (a duration of ca. 170 my) an extensive carbonate platform covered large parts of the Arabian craton (Glennie et al., 1974; Murris, 1980; Harris et al., 1984; Rabu, 1987; Scott, 1990; Pratt and Smewing, 1993; Alsharhan, 1995; van Buchem et al., 1996; van Buchem et al., 2002). During the Early Cretaceous Albian stage, several distinctive paleo-depositional belts shaped this platform. Alluvial to coastal deposits rimmed the Arabian-Nubian Shield in the southwest, grading north-eastwards into a shallow-marine, carbonate shelf (Figure 2). In Oman, this carbonate shelf was superposed by a depression, here referred to as the ‘Nahr Umr Basin’. North-westwards, the Nahr Umr Basin probably opened into the larger and deeper Bab intrashelf basin of the United Arab
Figure 1: Geotectonic map of Eastern Oman with indication of the (para)autochthonous sedimentary cover of Arabia. Refer to color code for the main lithologic units. North-south transect across different outcrops of the Albian Nahr Umr and Al Hassanat formations is indicated as red line. The subsurface Fahud Basin and South Oman Salt Basin are shown in pink and contain the infra-Cambrian Ara Salt.
Figure 2: Palinspastic reconstruction of the Arabian platform in the Albian. (a) Arabian platform with the exposed basement in the southwest and the shallow-marine shelf to the northeast. Note position of Figure 2b. (b) Enlargement of the Oman portion of the Arabian platform with indication of the Albian platform top to basin transect (A-A’). (c) Schematic bathymetric transect A-A’ with indication of the terminology used.
Basement highs bordered the Nahr Umr Basin perhaps along its eastern margin (Haushi-Huqf High; Hughes Clarke, 1988; Ries and Shackleton, 1990), and its northern margin (Masse et al., 1997; Immenhauser et al., 2001). Sedimentary units show regional thinning and environmental shallowing towards these structural highs.

During late Aptian to late Albian time, a major influx of terrigenous material from the emerged Arabian-Nubian Shield (Figure 2; Murris, 1980; Hughes Clarke, 1988), and possibly from the Haushi-Huqf High, led to the deposition of the argillaceous Nahr Umr Formation. Throughout the southern Gulf, the Nahr Umr is principally an impermeable unit of clayey facies acting as a seal for the lower Aptian Shu’aiba/Upper Thamama carbonate reservoir rocks (van Buchem et al., 2002). In Oman, the Nahr Umr forms part of the parautochthonous Wasia Group of the Hajar Supergroup (Pratt and Smewing, 1993; Rabu, 1987; Immenhauser et al., 2000a, b).

Rudist-coral-algal lithosomes of Aptian to Albian age (Al Hassanat Formation; Masse et al., 1997; Immenhauser et al., 2001) rimmed a shallow swell trending along the northern continental margin of Oman facing the Hawasina Sea (Figure 2; Immenhauser et al., 2001). The Al Hassanat Formation platform (slope, margin, and lagoon; Figure 2c) is thus a small feature superposed on the craton-scale platform that covered Arabia during the middle Cretaceous. In terms of platform physiography, the Al Hassanat Formation formed a wave-resistant margin both towards the Hawasina Sea to the north (exposed in Wadi el Assyi, Figure 1), and perhaps towards the Nahr Umr Basin to the south (non-exposed). Masse et al. (1997) argued that these margins were separated by an emerged basement high (‘structural high’ in Figure 2b). The areas to both sides of the emerged high and the respective platform margins are here referred to as ‘Al Hassanat Lagoon’ (Figure 2c). Similar bio-constructed buildups might have flourished on the Haushi-Huqf High to the east of the Nahr Umr Basin (Immenhauser et al., 1998), but were eroded at a later stage. Southwards, the deposits of the Al Hassanat Platform interfingered with the argillaceous deposits of the open and deep shelf, the Nahr Umr Basin (Figure 2c).

**ALBIAN SEA-LEVEL RECORD INoman**

The Albian relative sea-level record in Oman (Harris and Frost; 1984; Davies, 1984; Sharland et al., 2001) was investigated in six localities along a 500-km-transect extending from the northern slope of Jabal Akhdar to the southern Huqf area (Figures 1 to 3). 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; 2000b). 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 (Figures 3 and 4; Immenhauser and Scott, 2002). Refer to Immenhauser and Scott (2002) for more detail of sea-level reconstruction. The sea-level record from the following localities is used for this study.

**Albian Al Hassanat Formation Platform Top**

**Wadi el Assyi Section**

The Albian platform top is equivalent to the Nahr Umr Formation (the Al Hassanat Formation) and exposed in Wadi el Assyi near the village of Nakhl (Figure 1; Masse et al., 1997; Immenhauser et al., 2001). The facies is dominated by coral-algal-rudist lithosomes at the platform margin, and lagoonal deposits in the more proximal settings, i.e. towards the emerged basement (Figures 3 and 4). The measured section ranges from the uppermost Aptian to the lowermost upper Albian. In Wadi el Assyi, at least 29 exposure events are recorded of which discontinuity surfaces (DS) 2a, 3a, and 4a are related to high-amplitude sea-level change (Figures 3 and 4). These surfaces were correlated with their equivalent exposure surfaces in the adjacent Nahr Umr intrashelf basin to the south of Jabal Akhdar, a distance of about 45 km (Wadi Mu’aydim in Figures 1 and 3). Exposure surfaces 1b through 26b are believed to be the result of low-amplitude regressive events. This conclusion was drawn because they are not recorded in the bathymetrically deeper settings of the Nahr Umr intrashelf basin (Figures 3 and 4). In the outer platform margin, surfaces 2a is superposed on surface 3a due to platform top truncation.
Figures 4a and 4b: Relations between physiographic positions of sections, bathymetry and amplitudes of sea-level change. (a) Overview of bathymetric and sedimentologic features in each section that aided in the estimate of the hydrodynamic level and thus the relative water depths. (b) Conceptual sketch showing the relation between physiographic positions of sections and amplitudes of sea-level cycles.

Figure 3 (facing page): Stratigraphic columns of the Nahr Umr at the localities under consideration. The shallow platform is to the right (Wadi el Assy, cf. Figure 2) and columns to the left are from the intrashelf basin margin and basin. Note different scale of the Wadi Mu’aydim section. Discontinuity surfaces are indicated in red (high amplitude sea-level change), blue (intermediate amplitude sea-level change) and green (small amplitude sea-level change). Sequence stratigraphic interpretation is given to the right of each section. Note key to color code and symbols in the upper left corner. Black, horizontal lines to the left of the Wadi Bani Kharus section indicate inferred 4th and higher-order cycle tops.
during exposure. In the adjacent platform slope setting, both discontinuities are separated in space and time (Figure 5; Immenhauser and Scott, 2002). This interpretation represents a revision of previous work (Immenhauser et al., 2001), and resulted in a revised nomenclature of surfaces in Wadi el Assyi. Figure 5 compares the two interpretations.

Nahr Umr Intrashelf Basin Margin

Wadi Bani Kharus Section

The proximal intrashelf basin margin of the Nahr Umr Basin in Oman is well exposed in Wadi Bani Kharus (Figures 1, 3 and 4; Immenhauser et al., 1999, 2000a). The facies is primarily dominated by argillaceous Orbitolina packstones and three calcareous marker bed successions of which the uppermost (Marker Bed Succession III - MBS III; Figure 3) is a rudist floatstone. The measured section, from the top of the underlying Shu’aiba Formation to the base of the overlying Natih Formation is latest Aptian to latest Albian in age. In Wadi Bani Kharus, ten regressive-transgressive cycles of intermediate to high amplitude are recorded (Figure 3). Nine of these discontinuities are correlated in time to an adjacent section about 2.5 km west of Wadi Bani Kharus. Discontinuity 7a apparently is local in extent, and could not be correlated in other nearby sections (Figure 3).

West of Wadi Bani Kharus Section

The section west of Wadi Bani Kharus (Figures 1, 3 and 4; Immenhauser et al., 1999, 2000a) is measured across the northern dip slope of Jabal Akhdar. The section is directly comparable in age, facies and physiographic location to the section in Wadi Bani Kharus. This locality was chosen to capture lateral facies variability in the spatial domain of a few kilometers. Ten discontinuity surfaces reflect relative sea-level fall of intermediate to high amplitude and thus transient exposure of the Albian carbonate seafloor at this locality. One of these surfaces is apparently local in origin (Figure 3; surface between 9a and 10a). Six discontinuities are correlated in time to the more proximal intrashelf basin margin as exposed in Wadi Mu’aydim, a distance of about 45 km (Figures 1, 3 and 4).

Wadi Mu’aydim Section

The deeper intrashelf basin margin of the Nahr Umr Basin in Oman is exposed in Wadi Mu’aydim at the southern slope of Jabal Akhdar (Figures 1, 3 and 4; Simmons and Hart, 1987; Scott, 1990; Immenhauser et al., 1999). The facies is in principle comparable to Wadi Bani Kharus but overall the carbonate content is lower. As in Wadi Bani Kharus, the MBS III is a rudist floatstone (Figure 3). In Wadi Mu’aydim, the Nahr Umr (and under and overlying formations) is stratigraphically much thicker (235 m) compared to the Nahr Umr Formation in Wadi Bani Kharus (109 m). This points to an higher average rate of accommodation space generation (subsidence rate) in this area. Based on biostratigraphy and the correlation of marker surfaces and marker bed successions, the age of the Nahr Umr Formation in Wadi Mu’aydim is latest Aptian to latest Albian (Immenhauser et al., 1999). Six subaerial exposure surfaces record sea-level change of intermediate to high amplitude, and four of these discontinuities are correlated across a distance of about 120 km to the intrashelf basinal settings recorded at Jabal Madar (Figures 1, 3 and 4).

Jabal Madar Section

The Nahr Umr Formation at Jabal Madar, situated at the easternmost end of the Adam Foothills (Figure 1), represents the Albian intrashelf basin setting in Oman (Figure 2; Immenhauser et al., 1999). The Albian facies at Jabal Madar is barren and argillaceous and only the MBS I-III protrude as calcareous units (Figure 3). Based on biostratigraphy and the correlation of marker bed successions (MBS in Figure 3), the stratigraphic range of the Nahr Umr at this locality is comparable to the more proximal sections (uppermost Aptian to uppermost Albian; Immenhauser et al., 1999). Four high-amplitude regressive events and subsequent transgressions are recorded in discontinuity surfaces and overlying argillaceous
Immenhauser and Matthews
Albian sea-level cycles in Oman: the ‘Rosetta Stone’ approach

Figure 5: Key to discontinuity surfaces in Wadi el Assyi. Note changes between (a) Immenhauser et al. (2001) and (b) Immenhauser and Scott (2002), which is used in this paper. Schematic model for the formation of surfaces 2a and 3a is given to the right (c). Discontinuity 2a formed during sea-level fall that emerged the platform margin and the upper slope of the Al Hassanat platform in Wadi el Assyi (t2). Subsequently, the platform top was flooded again during t3. A renewed sea-level fall (t4) truncated the platform top and DS 3a was superimposed on DS 2a. The two surfaces are spatially resolved on the upper slope but merged at the locality of the Wadi el Assyi section. This later observation required a reinterpretation of the surfaces in Wadi el Assyi.
deposits. Surfaces 4a, 6, and 8a cap the calcareous MBS I-III (Figure 3). Each of these marker bed successions represents a period of forced regression of the platform top carbonate factory during periods of sea-level fall (Immenhauser and Scott, 2002). All four discontinuities are correlated southward with their equivalents in the Nahr Umr Formation in the southern Huqf, a distance of about 300 km. Industry-generated gamma ray data across the Nahr Umr Formation are available from the nearby Jabal Madar exploration well.

**Southern Huqf Section**

The stratigraphically thinnest (34 m), and perhaps the most basinal section of the Nahr Umr Formation along this transect, is exposed in the southern Huqf, to the west of the village of Duqm (Figures 1, 3 and 4; Dubreuilh et al., 1992; Immenhauser et al., 1999). As at Jabal Madar, the facies is argillaceous with the MBS I-III protruding as calcareous units (Figure 3). Biostratigraphic data, particularly the orbitolinid fauna, from this section suggests that the stratigraphic range is comparable to that of the other Nahr Umr sections. This is confirmed by the presence of MBS I-III, which are, generally speaking, relative time-markers. Here, as at Jabal Madar, high-amplitude sea-level drop resulted in the formation of discontinuity surfaces. It is, however, uncertain whether at this locality sea-level fall resulted in terrestrial subaerial exposure of the carbonate seafloor as evidence for emergence is very poor. If not, then sea-level fall was at least sufficient to lower the effective wave base onto the seafloor, a process that resulted in winnowing of carbonate sediments and the formation of prominent manganese-stained marine hardgrounds (Immenhauser et al., 2004).

**Estimates of Albian Paleo-Water Depths**

Estimates of the average paleo-water depths at the different localities investigated are crucial inasmuch as they provide the basis for an assessment of the magnitudes of Albian sea-level change. The approach applied here involves two steps.

The first step is based on field and thin section petrographic evidence (facies and sedimentary features) pointing to specific hydrodynamic levels, which in turn, are related to specific bathymetric domains. In a simplifying manner, Albian sedimentary rocks in Oman fall in three categories, each indicative of a specific hydrodynamic level: (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). See Table 1; Immenhauser et al. (1999) and Immenhauser and Scott (2002) for more detail.

The relative bathymetric positions of the six Nahr Umr sections is laid out below and illustrated in Figure 4a. The shelf-margin reefal and lagoonal carbonates that build the Wadi el Assyi section were deposited above the fair-weather wave base. By analogy to modern carbonate platforms, we assume that this took place in water depths of between 10 and 30 m. The carbonates that build the Wadi Bani Kharus section and the section west of this locality (Figure 1) are characteristic for a below-fair-weather wave setting, but numerous tempestites show field evidence for transient storm waves and currents (Immenhauser et al., 1999). The same setting, but perhaps more basinwards, is reflected in the more argillaceous Wadi Mu’aydim section. The Jabal Madar and Southern Huqf sections are largely built by argillaceous sediments (exceptions are the calcareous marker bed successions; Figure 3) that show no evidence for any hydrodynamic agitation and tempestites are lacking. It is thus assumed that these rocks were deposited beneath the reach of storm waves or currents.

Step 2 uses recent analogs (i.e. the Arabian Gulf and the Gulf of Carpentaria) in order to asses the depth of the fair weather and storm wave base in the Albian Nahr Umr Basin of Oman. As discussed in Immenhauser and Scott (2002), the present fair-weather wave base in the Gulf of Carpentaria is at about 35 m, whereas the effective storm wave base in the Arabian Gulf is at about 40 m depth near islands and at about 70 m on the open ramp (see references in Immenhauser and Scott 2002). Assuming that the these recent analogs provide wave-base depth estimates that are in the correct order of magnitude, we place the Albian fair-weather wave base in the Nahr Umr Basin at about 30 m and the storm wave-base at about 50 m, respectively (Figure 4). The implication is that Albian sea-level fall that exposed the carbonate seafloor at Jabal Madar had amplitudes of several tens of meters (Figure 6).
ASSESSMENT OF RATES OF SEA-LEVEL CHANGE

Each regressive-transgressive (sea-level) cycle can be described with two basic characteristics:

1. The amplitude (magnitude) of sea-level fall and subsequent rise, here expressed in meters; and
2. the period, i.e. the frequency with which sea-level cycles occur in time.

This is exemplified in the case of the high-amplitude regressive event that is recorded in discontinuity surface 4a (DS 4a; Figures 3, 6) in all Nahr Umr sections. The inferred amplitude of sea-level fall that resulted in the formation of this surface is on the order of 50 m. Biostratigraphic data (Immenhauser et al., 1999) suggest that large-scale discontinuity surfaces (DS 4a, 6a, 8a; Figure 3) have a periodicity on the order of about 1.5 to 3.5 my (Figure 6). From this amplitude and the periodicity the resulting rate of sea-level fall and rise is estimated to be on the order of 10s of m/my (see discussion in Immenhauser and Scott, 2002).

In the case of small-amplitude, regressive-transgressive cycles in the platform section in Wadi el Assy (Figures 3 and 6), here the estimated amplitude is about 10 m. Based on biostratigraphic data again, the periodicity of, for instance, DS 9b though 16b, is in the order of 200 to 400 ky (Figure 6). This results in a rate of sea-level change that is in the same order of magnitude (i.e. 10s of m/my), as the one deduced from discontinuity 4a resulting from a sea-level fall with a much higher amplitude (Figures 3 and 4b). Below, we evaluate the hypothesis that these characteristics (amplitude and period) are a key for our understanding of factors that caused the relative fluctuations in Albian sea level as recorded in Oman.
SEA-LEVEL FLUCTUATIONS – DRIVING MECHANISMS

In a general approximation, relative fluctuations in sea level are driven by either changes in the water volume in the world's oceans, or by changes in the shape of oceanic basins. The processes involved have been extensively discussed in the literature (see for instance Lambeck, 1988; Harrison, 1990; Tooley, 1993 for a review). In Figure 7, the most significant factors that influence the position of relative sea level are shown graphically. Each of these processes drives sea-level fluctuations with (within a certain range) characteristic amplitudes, periods and rates. Given the inferred frequency of the Albian sea-level events in Oman (100-1,000 ky), transient short-lived events such as Tsunami waves, storm surges, etc. are not further considered here. In contrast, mechanisms that cause sea-level change within the amplitude, period and rate domain of the Albian sea-level events in Oman (Figure 7) are discussed below.

**Changes in Ocean Basin Volumes**

**Geoid Deformation**

The concept of geoid-eustasy as a driving mechanism for Cretaceous sea-level fluctuations has been proposed by Mörner (1981). The term ‘geoid’ refers to the irregular surface of the Earth (sea and land) characterized by humps and depressions with a presently maximum vertical difference of 180 m (Mörner, 1976). Any change in a factor controlling the rotation of the Earth, or the distribution of mass (gravity), will affect the geoid configuration expressed as geoid-eustasy. The basic principles of geoid deformation with time, and its affects on the relative position of sea-level, are now generally accepted. Considerable disagreement, however, exists on the rate and period of sea-level change driven by these processes. Mörner (1980) suggested that cyclical sedimentation patterns (periods of 80-90 ky) in lower Albian sediments from the Bay of Biscay were controlled by geoid-eustasy. Current opinion seems to be that the present-day geoid represents events on the Earth that happened many millions of years ago (see discussion in Harrison, 1990). This implies that changes of the geoid occur very slowly (Lambeck, 1988). Dewey and Pitman (1998) estimated that geoid-related, sea-level changes have rates in the order of about 4 m/my. Too slow thus, to explain the regressive-transgressive cycles in the Albian of Oman.

**Volume Changes of Mid-ocean Ridge Systems and Spreading Rates**

It has been argued (but disputed, cf. Heller et al., 1996) that during the Albian, the ocean-crust production rates were exceptionally high (Larson and Kincaid, 1996). Kaiho and Saito (1994) estimated production...
Figure 7: Overview of processes that cause relative sea-level change with indication of their spatial (linear) dimensions and their magnitude plotted against time. Note that estimates of dimensions, durations and magnitudes of sea-level change are tentative approximations and differ depending on the source used (see text for references). The range of Albian sea-level cycles in Oman are shown as red box. Stress-related deformation of margins (4b) has an insufficient spatial dimension (100-200 km) and ridge volume changes (2) cause lower amplitude and lower frequency sea-level changes than those reconstructed from Oman. The best fit between the Oman sea-level spatial dimensions, durations, and magnitude is with 8 (glacio-eustasy; Figure 7d). (a) Plot of spatial dimensions of geoid, tectonics and volcanism-related relative sea-level change versus duration, and (b) plot of magnitude of geoid, tectonics and volcanism-related sea-level change versus duration must be viewed as pairs. The same accounts for (c) Plot of spatial dimensions of glacial and steric (volume changes of seawater due to salinity and temperature changes) effects, sediment loading of oceanic basins, and flooding of low-lying basins versus duration, and (d) plot of magnitude of glacial and steric (volume changes of seawater due to salinity and temperature changes) effects, sediment loading in oceanic basins, and flooding of low-lying basins versus duration.
rates that range from 48 to 53 x 10^6 km^3/m/yr. The underlying mechanisms of these unusually high spreading rates between 124 and 83 Ma are summarized in the concept of ‘superplumes’ (Larson, 1991). This term refers to extraordinary upwelling of heat and deep-mantle material in the form of one or several very large plumes. These superplumes coincide with increases in world temperature, deposition of black shales, oil generation, and a high global sea level.

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 one centimeter per ten year increase in the spreading rate is 54 m. The consequent rise is most rapid immediately after the change, and decreases exponentially with time. Only the most extreme cases of spreading change, however, cause sea level rise with a rate of 10 m/yr (Dewey and Pitman, 1998). The implication is that changes in spreading rates were probably an important factor influencing mid-Cretaceous sea level, but one that had lower amplitudes and lower frequencies (Komintz, 1984) than the sea-level oscillations recorded in Oman (Figure 7a, b).

**Hot Spots, Trap Basalts, and Seamount Volcanism**

McNutt et al. (1990) proposed that, at approximately the Aptian-Albian boundary, the northern Darwin Rise (Western Pacific) was elevated due to intense volcanic activity. The estimated magnitude of sea-level rise that resulted from this uplift is of the order of about 30 m (Dewey and Pitman, 1998). Other studies that compiled data on the volcanogenic sediment distribution in the Indian Ocean found high abundances of volcanogenic components in Albian marine deposits pointing to intense volcanic activity (Kerguelen hotspot, Rajmahal Traps in northeast India, volcanic activity on the Ninety East Ridge, and the Broken Ridge) throughout the Albian (Sykes and Kidd, 1994). These observations highlight the significance of changes in the volume of Albian ocean basins due to volcanic activity. The magnitude of the estimated sea-level rise related to Albian thermal activity is in the order of 30 m for the period 110 to 70 Ma; comparable to the amplitude of Albian sea-level cycles in Oman. The rates, however, of hotspot driven sea-level rise are slow (0.8 m/yr; Dewey and Pitman, 1998) when compared to the rates of sea-level change in Oman (10s of m/yr). This implies that seamount volcanism cannot explain the regressive-transgressive events recorded in the Nahr Umr Formation (Figure 7a, b).

**Localized Block Movement, Regional to Plate-scale Tectonism and Plate Margin Deformation**

Given the semi-regional extent of the sea-level events in Oman (approximately 500 km), localized block scale (one to 10s of km) tectonic movements that might have caused local relative changes in sea level are not considered here. Likewise, we do not consider plate-scale collisional (orogeny) events that might influence the shape of adjacent oceanic basins. These collisional events cause slow sea-level change with rates in the order of 0.65 to 0.88 m/yr (Dewey and Pitman, 1998), and cannot explain the Oman Albian regressive-transgressive cycles. These mechanisms are thus not further considered here (Figure 7a, b).

The deflection of the lithosphere at passive continental margins is dominated by sediment loading and thermal contraction (Cloetingh, 1986). Based on this observation, Cloetingh et al. (1985) proposed a tectonic mechanism for regional relative sea-level variations with rates of 1-10 m/yr and magnitudes of up to a few hundred meters. The model explains these sea-level changes, provided that horizontal stress fields occur at continental margins on geologic time scales. The model represents the interaction between these stresses and the deflections of the lithosphere caused by sediment loading. If regional stress reorganizations at continental margins occur at the time scale of one million years, then sea-level changes in the order of the major Albian cycles observed in Oman are possible. The actual magnitude obtained for a given change in stress is controlled by the magnitude of the perturbation or deflection of the lithosphere at the time when the in-plane stress is applied. Possible perturbations of the passive plate-margin stress field are controlled by the sedimentation rate at this margin and, related to this, the response of the lithosphere to this sediment load (Cloetingh, 1986).

Depending on the position of a hypothetical ‘observer’ at this margin, sediment and water loading will cause basement subsidence at one locality and asthenosphere rise will cause basement uplift at another locality. Both processes cause unidirectional changes in relative water depth (i.e. deepening
or shoaling) as opposed to the cyclical regression-transgression events as observed in interior Oman. Furthermore, the Albian Oman sea-level fluctuations are observed across a distance of at least 500 km perpendicular to the northern Oman margin (Figure 1). In contrast, plate-margin deflection is a process that is expected to affect not more than a 200-250 km wide zone of the plate margin (Ravaut et al., 1993). This implies that the passive margin deflection mechanism is an unlikely explanation for the Albian sea-level fluctuations in Oman (Figure 7a, b).

**Catastrophic Flooding of Areas Below Sea Level**

The near instantaneous marine flooding of depressions such as the present-day Death Valley and the Dead Sea Valley (lying below the sea-level) is a feature that is known from the Holocene Black Sea, Messinian Mediterranean Sea, and Aptian South Atlantic (Dewey and Pitman, 1998). The flooding of these basins results in a global sea-level fall, at very high rates. Estimates of the amplitudes of sea-level fall range from about 10 m in the case of the Messinian Mediterranean, to about 0.1 m in the case of the Holocene Black Sea. As far as it is known, there is no geologic evidence for comparable and repeated desiccation-flooding cycles of low-lying Albian depressions. This mechanism is thus not further considered here (Figure 7c, d).

**Ocean Sediment Volumes**

Harrison et al. (1981) discussed the possibility that changes in the amount of sediment deposition in the major oceanic basins could have a significant effect on sea level via a change in ocean basin volume. The average thickness of carbonates in the ocean basins today is around 300 m, 90% of these sediments are composed by the tests of pelagic foraminifera, a life form that did not evolve into volumetrically important sources of deep-sea sediments prior to the Late Cretaceous. Other sources of large sediment volumes are young rising orogens such as presently the Himalayas. It is obvious, however, that the long-term evolutionary trends of planktonic organisms, or the erosion products of young rising mountain chains cause sea-level changes that are of much longer periods than the regressive-transgressive cycles observed in Oman. Dewey and Pitman (1998) estimated that changes in sediment input can cause sea-level change in the order of 0.6 to 2.7 m/my. Therefore, we do not further discuss these processes here (Figure 7c, d).

**Changes of Ocean-water Volumes**

**Steric Effects**

Steric effects refer to changes in seawater volume associated with changes in the density of seawater (temperature and salinity). At its present salinity (~ 35‰), seawater has a maximum density at close to 0.0°C. Thus, as seawater temperature increases, density decreases and seawater expands (Wigley and Raper, 1993). The amount of expansion per unit rise in temperature is determined by the thermal expansion coefficient (a function of temperature, salinity and pressure, e.g., Bryan and Cox, 1972). In tropical latitudes the present expansion coefficient is around 297 x 10^-6 per degree Celsius at 25°C.

Assuming steric seawater expansion is the only active factor influencing the position of the sea surface, then sea-level change in the order of several 10s of meters would require repeated (and rapid) seawater warming and cooling in the order of several tens of degrees. This requirement is not in agreement with estimates of Cretaceous seawater temperatures (Frakes, 1999; Frakes and Francis, 1988). Accordingly, steric effects as the driving mechanism of Albian sea-level fluctuations in Oman is not further considered here (Figure 7c, d).

**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 speculated about the possible presence of continental ice shields in the Cretaceous Period (see Francis and Frakes, 1993; Stoll and Schrag, 1996; or Frakes, 1999 for an overview). However, the modeling of Antarctic glaciers (e.g. Fastook and Prentice, 1994;
discussed briefly in Prentice and Matthews, 1991) gives some idea of what not to expect around the perimeter of even a fairly large ice sheet situated in proximity to the South Pole.

Such ice sheets are fed almost perpetually with snow by the general circulation. Descending air of the polar vortex provides a continual fall of “fair weather” light snow over the entire ice sheet. Thus there is vertical accumulation of ice over the extent of the ice sheet, leading to constant ice creep to lower topography and lower latitude. In contrast, melting occurs as the seasonal interaction of atmosphere and ice at the perimeter of the ice sheet. Because of the topography and temperature of the ice, such interaction extends no more than a few hundred kilometers inwards from the ice sheet margins, creating a vast braided stream terrain, even in close proximity to the ice sheet.

During the Albian, Antarctica and Australia were positioned at mid to high latitudes south (Figure 8). Central Australia was occupied by the large intracratonic Eromanga Basin, in which coastal sand and marine marls were deposited (Frakes and Francis, 1988). The marine marls contain exotic clasts
of mainly Precambrian quartzite and volcanic rock up to 3 m in diameter. Hydrodynamic boundary conditions suggest that these exotic blocks were transported by floating ice and dropped after melting of the ice (Frakes and Francis, 1988). Thus, whereas skeptics may attribute drop stones to seasonal freezing of shoreline gravels (although exotic blocks with 2-3 m in diameter can hardly account as ‘shoreline gravel’; Frakes et al., 1995), the important question becomes whether the geologist can distinguish tillites from braided stream deposits. Tillites implies ice sheet or mountain glacier (Figure 9). Thus, paleogeographic context must be considered. Deposits such as till have a low preservation potential (Price, 1999). For example, most sediments of glacial origin are transported to the marine environment and there, they are likely to be completely eroded thereby eradicating any evidence of glacial origin (Eyles, 1993). We thus acknowledge that the glacio-eustasy conclusion (even as qualified below) is controversial due to the absence of unequivocal evidence for major perennial ice sheets (ice bergs as opposed to winter shoreline ice; see Price, 1999 for a detailed discussion).

The field area most likely to prove or disprove the existence of Cretaceous ice sheets is presently beneath three kilometers of East Antarctic ice. 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).

Another point that merits consideration is the relation between relief and average temperature. Given the fact that the Transantarctic Mountains stretch across 3,500 km, a relief difference of 1 or 2 km would allow for the existence of seasonally moderate climates in lower latitudes (i.e. intermontane basins and coastal plains covered with forests), contrasted by much colder average temperatures (and thus below freezing temperatures) at higher elevations where substantial perennial ice sheets could exist. Indeed, during the Albian-Cenomanian gymnosperm rainforests first appeared in Antarctica but Aptian-Albian floral information from Victoria, Australia was interpreted as indicating cool and wet temperate climates (Douglas and Williams, 1982). Refer to Price (1999 and references therein) for an extensive discussion of this matter.

In summary, it thus seems that sea-level cycles in Oman either reflect un-named processes we do not understand, or they reflect changes in continental ice volumes (glacio-eustasy). This conclusion is particularly controversial due to the ongoing debate centering on the presence or absence of major continental ice shields during the Cretaceous (Weissert and Lini, 1991; Francis and Frakes, 1993; Sellwood et al., 1994; Price, 1999). To more quantitatively evaluate the glacio-eustasy hypothesis we have undertaken modeling studies. First, we evaluate the sensitivity of sea-level calculations to modification of boundary conditions. Then, we construct simulated stratigraphic sections using the new sea-level calculations.
BIOSTRATIGRAPHY, CHRONOSTRATIGRAPHY, ORBITAL FORCING TIME SCALES AND RELATIVE TIME CORRELATION

Biostratigraphic Time Scale

The dating of the Nahr Umr sections is based on biostratigraphy, mainly orbitolinid foraminifera and red algae, resulting in relative biostratigraphic ages whereas the Al Hassanat (platform section) is mainly dated by macrofossils (e.g. rudist bivalves) in combination with microfossils (Masse et al., 1997; Immenhauser et al., 2001). Although the age-diagnostic biota in these sections is not diverse, it is typical of the Albian of Oman. The error bar related to biostratigraphic age resolution in these Oman Albian shallow-water sections is on the order of 1.0 my and becomes larger with an increasingly impoverished biota in more restricted environments.

In order to improve this relatively coarse biostratigraphic time framework the technique of graphic correlation (e.g. Shaw, 1964; Carney and Pierce, 1995) was applied to date the Albian sections. For this purpose, the mid-Cretaceous composite standard was used (MID CS; Scott, 1990). MID CS is a database of more than 40 sections located in the Tethyan Realm, in Europe, Africa, the Middle East, and North America. Stage boundaries are defined by European reference sections that have been graphed into the database and calibrated to the Harland et al. (1990) time scale. Nevertheless, more recently, new chronostratigraphic ages for the Aptian-Albian and Albian-Cenomanian boundaries have been proposed (Gradstein et al., 1994). According to Gradstein et al. (1994), the error bar for the Aptian/Albian boundary is 1.1 my, and for the Albian/Cenomanian boundary 0.6 my. These error bars are sufficient to include the Harland et al. (1990) time scale with respect to the Aptian/Albian boundary (Harland et al., 1990 = 112 Ma; Gradstein et al., 1994 = 112.2 ± 1.1 Ma). However, for the Albian/Cenomanian boundary the two time scales differ by 1.9 my i.e. more than the error bar of 0.6 my assigned by Gradstein et al. (1994). There are also substantial differences when comparing the Albian time scale of other authors (e.g. 5 my for the Aptian/Albian boundary when comparing Haq et al., 1987 with Gradstein et al., 1995). This must be kept in mind when assessing the Albian chronostratigraphy of Oman.

Based on the MID CS, and calibrated against the Harland et al. (1990) time scale, the following age estimates were made for the discontinuity surfaces (DS) in Wadi Bani Kharus (Figure 3) and extrapolated into other sections. Discontinuity surfaces 1a ~ 113.8 Ma; DS 2a ~ 112.5 Ma; DS 3a ~ 111.2 Ma; DS 4a ~ 108.2 Ma; DS 5a ~ 104.8 Ma; DS 6a ~ 103.2 Ma; DS 7a ~ 103.1 Ma; DS 8a ~ 101.8 Ma; DS 9a ~ 101.6 Ma; DS 10a ~ 99.9 Ma (Immenhauser et al., 1999). Graphic correlation assigns no specific error bars to these ages but instead they represent the geologically most reasonable model age correlation as suggested by the MID CS. It is implicit that the error bars would be at least as large as those assigned to the time scale in general.

Orbital Forcing Time Scale and Cycle Orders

Orbital forcing identifies numerous higher-order cycles (eccentricity: 100 ky; tilt: 40 and 54 ky; and climatic precession: 19 and 23 ky). Orbital forcing time scales do not make reference to biostratigraphy. Rather, they are calculated from the present-day back through time from fundamental frequencies for the motion of each planet about the Sun, and an initial phase angle for each term (Matthews and Frohlich, 2002). The trigonometric terms and their initial phase angles are derived by Fourier analysis (e.g. Berger and Loutre 1991; Berger et al., 1992) from the numerical integration of the gravitational effect of each planet on all other planets, back through time (e.g. Laskar, 1990; Quinn et al., 1991).

A corollary of the original Graphic Correlation concept (e.g. Shaw, 1964) was that geologists and paleontologists could make large progress on the fundamental question, “Can we correlate?” without reference to absolute time. To that, orbital forcing adds the expectation that 404-ky (4th-order cycles) and 2.4 ± 0.4 my (3rd-order cycles) will be manifested so long as the Earth has ice sheets on continents. These two periods are stable and expected on relative time scales from Precambrian to Recent.

Orbital Forcing Cycle Orders

The principle advantage of orbital forcing time scales is the identification 404 ky and 2.4 my modulation as a major pattern of climate oscillations and thus sequence stratigraphy (e.g. Matthews and Frohlich,
2002; their figures 8, 9, 11 and 12). From the modeling perspective, 4th-order cycles are 404 ky (not an arbitrary 100 ky to 1.0 my) and identify this as the ‘tuning fork’ of geologic time (Matthews and Frohlich; 2002). Likewise, 3rd-order cycles are 2.4 ± 0.4 my (not an arbitrary 1.0 to 10 my).

By this calculation, rates of sea-level change for higher-order cycles are the same as for nearby 4th-order cycles; thus the shorter the period, the smaller the amplitude. However, on the 100 my time-scales, the shorter the period, the more likely that phase relations to longer period terms are in error. Further, Matthews and Frohlich (2002) found these discrete periods could not be consistently recognized in these sea-level constructs. They suggest leaving all periodicities above 4th-order as a single category as ‘higher-order cycles’.

We expect each 3rd-order cycle to consist of a number of 4th (and higher)-order cycles. The modeling suggests this number to vary, especially with regards the number of 4th-order cycles per 3rd-order cycle. From a modeling perspective, two clusters are predicted. Third-order cycles of 2.4 ± 0.4 my translate to containing five to seven 4th-order cycles. On occasion, what is a 3rd-order node in the sense of eccentricity (Matthews and Frohlich, 2002; their figure 8) goes understated in the sea-level calculation, thus creating a predicted double cycle of 4.8 ± 0.4 my. This translates to a single observable cycle that would contain eleven to thirteen 4th-order cycles. Importantly, one of the simplest predictions of orbital forcing is that there should be a sort of ‘Morse Code’ of short (3rd-order) and long (two 3rd-order) cycles in the observed record.

A caveat on this remains that as geologic time increases, the higher frequencies will be the first component to become imprecise. The higher the frequency, the sooner phase relations will be expected to deviate from prediction. Thus, for a time interval as old as is under consideration here, precision on higher-order frequencies does not allow prediction of which 4th-order cycles will be simple and which complex. Thus, it is easy to see how the observer might over-count presumed 4th-order cycles. Nevertheless, an under-count would suggest a shortcoming of the recording system or a problem with the model.

**Orbital Forcing Time**

Given the hypothesis that 2.4 ± .4 my, 3rd-order cycles (e.g. Matthews and Frohlich, 2002; their figure 8) are fundamental, they can be calculated back to, for example, 200 Ma. Sensitivity tests produce sizeable differences in absolute age on the 100 my absolute time scale, however, the signature of the individual 3rd-order cycles remains largely intact. Thus, on a relative time scale, the duration of the Albian (assuming it can be defined in sequence stratigraphic terms) may be a given duration ± 0.4 my, but the absolute age of any event might be a given age ± 3 my. To clarify the distinction between absolute time and orbital-forcing time, we refer to the latter as ‘Ma (OFT)’. Further cooperation between stratigraphers and modelers is required to improve existing differences between these approaches and to converge towards a refined data-model fit.

**ORBITAL FORCING OF SEA LEVEL**

We follow the simple ‘parametric forward modeling’ (PFM) strategy of Matthews and Frohlich (2002, p. 515-520). As indicated in Figure 10, first, we use a PFM (IceCylinder 2.91, Matthews and Frohlich, 2002, p. 516), which inputs orbital forcing time series and outputs a calculated sea-level time

---

**Steps in Parametric Forward Modeling (PFM) Used Here**

| Orbital forcing | Ice Cylinder | Sea level | SimStrat | Stratigraphic column |
|-----------------|--------------|-----------|----------|---------------------|

Figure 10: Sketch illustrating steps in parametric forward modeling (PFM) approach used here.
series. Second, we use the model software SimStrat, which inputs sea-level time series and outputs a simulated stratigraphic section. By appropriate choice of geologically reasonable parameters, the output is designed to produce a simulated stratigraphic section that resembles any explicit observed stratigraphic section.

Matthews and Frohlich (2002) sea-level calculation ‘A’ is a general-purpose calculation intended to evaluate the general tendency of sea-level response to orbital forcing. Their sea-level calculation ‘B’ was specifically calibrated to yield an acceptable shape comparison with Haq and Van Eysinga (1987) and Haq et al. (1987) relative sea-level curve for portions of the Jurassic. Thus, we begin our quest for a suitable sea-level file by making small modifications to calculation ‘A’.

Calculation and Parameter Selection

The resulting sea-level calculation is presented in Figure 11. Beginning at 121 Ma (OFT), we made two changes to calculation ‘A’. The resulting calculation is henceforth referred to as calculation ‘A.2’. Our first change was to relax allowable minimum sea level from -50 m to -90 m relative to an ice-free world. Technically, this volume of ice might slightly exceed the present capacity of Antarctica. However, this model is not capable of duplicating the entire dynamic range from nonexistent ice sheet to maximum ice sheet on Antarctica (Figure 8). The dynamic range presented in calculation ‘A.2’ is approximately 60 m sea-level equivalent; a value quite compatible with the scale of Antarctica.

Our second change to the calculation was to speed up the entire ice-making/ice-melting system. For simplification of the model, Matthews and Frohlich (2002) linked all ice-making or ice-melting terms to each other (i.e. all maximum-melt rate parameters are the same; all maximum-make rate parameters the same and 10% of maximum-melt rates). We left this arrangement untouched and specified an increase in the maximum rate of ice melting from 0.5000 to 0.6150 m/my. The PFM applies this parameterization (see Matthews and Frohlich, 2002; their figure 9) 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. All parameters relating to description of the ice sheet remain as specified for calculation ‘A’ (Matthews and Frohlich, 2002). The results of calculation ‘A.2’ are depicted in Figure 11 for 121 Ma to 98.4 Ma (OFT). Younger than 98.4 Ma (OFT), an additional small modification was made.

For the parameterization described above, an instability is encountered at around 93 Ma (OFT). Instabilities are common in this sort of modeling and in nature. Here, if a certain parameterization value is exceeded, the system will lock onto maximum ice volume and never recover. Or, alternatively, if a certain minimum size ice sheet is ever calculated, the model will lock onto minimum ice volume and never recover.

With regards to the 93 Ma (OFT) instability, the system locks in on maximum ice volume and varies slightly from that condition for the next several tens of millions of years of the calculation. In their sea-level calculation B, Matthews and Frohlich (2002) found similar instabilities that corrected themselves as the calculation went younger. We ran calculation ‘A.2’ up to 77 Ma (OFT) and it did not recover from small variation from maximum ice volume.

We experimented with several parameters to relieve the instability at 93 Ma (OFT). We began our modified calculation at 98.400 Ma (OFT) in order to see if changes in model parameterization changed the sea-level file 98.4-95.0 Ma (OFT) (changes are very small). We ultimately settled on modification to the parameter $I_{s,crit}$ as best exemplifying the concept of an instability. In physical terms, this parameter specifies the insolation value for Antarctica above which summer ice melting occurs, and below which ice making shall continue through the summer. Whereas this number is specified as 435 watts/sq m for calculation ‘A’, we found that changing the number to 434.8 was more than adequate to relieve the instability.

To demonstrate the sensitivity of such instabilities, we determined the number required to relieve the instability to great precision. One digit in the seventh decimal place is the difference between a
calculation locked in on high-frequency variation at and around maximum ice volume, and a calculation that does not quite touch maximum ice volume, and then goes on for tens of millions of years without even coming close to maximum ice volume.

The above parameterization arrives at an acceptable solution to the forward problem posed by the evidence for large, rapid sea-level fluctuations around discontinuity 6a in the Jabal Madar (Figures 1 and 3) stratigraphic section. Our sea-level calculation from 121 Ma to 85 Ma (OFT) is presented in Figure 11. The most striking feature of Figure 11 is the ubiquitous 404-ky cycle.

The second most striking feature of Figure 11 is that the 4th-order cycles are indeed bundled into quasi-periodic 3rd-order cycles. In general terms, there is a low sea level between virtually every 4th-order highstand. Third-order lowstands are more pronounced and commonly involve several 4th-order cycles (e.g. 100-101, 105-106 Ma (OFT)).

**COMPARISON OF SEA LEVEL CALCULATION A.2 TO OMAN ALBIAN STRATIGRAPHIC SECTIONS**

We identify 3rd-order lowstand 43.0 as the most likely event to correlate with Albian DS 6a in the Jabal Madar section of Oman (Figures 3 and 11). This expectation proved valid. The predicted abrupt sea-level lowering beginning around 95 Ma (OFT) presents a good fit to the onset of the deposition of the limestones that build the overlying Natih Formation (Figure 11), pointing to an overall shallower depositional environment.

![Figure 11: Calculated orbital forcing sea level file 'A.2'. Time units are millions of years orbital forcing time (OFT) and are herein regarded as relative time units (cf., Shaw, 1964). Numbers 41.0 through 51.0 are Matthews and Frohlich (2002) third-order cycle boundaries. These are correlated to Oman sequence boundaries 3a through 6a. See Figure 3.](http://pubs.geoscienceworld.org/geoarabia/article-pdf/9/3/11/5441661/immenhauser.pdf)
The reef margin and lagoon portions of the Wadi el Assy section (Figures 3, 6 and 11) fit well to a general orbital-forcing scheme. Spacing of DS 3a through 13b (Figure 11) is compatible with orbital forcing 4th-order cycles with higher-order intermittently represented (e.g. DS 9b, 10b). High energy, shallow-water facies occur during each 4th-order highstand, making any 3rd-order cycle hard to distinguish from an unending stream of 4th-order cycles.

In contrast, upon entering the Wadi el Assy lagoon section (here taken nominally to begin above DS 13b in Figure 3 (cf. Immenhauser et al., 2001), the facies are progressively lower energy, deeper water and two intervals without discontinuity surfaces (98-125 m and 139-170 m; Figure 3) are set off by intervals with abundant discontinuity surfaces (92-98 m, 125-139 m and 170-180 m; Figure 3). This is as to be expected in orbital forcing 3rd-order cycles in deeper water sections. In such section, several 4th-order cycles occur without subaerial exposure. Higher-energy facies would be interpreted as the record of orbital forcing 4th-order lowstands. Indeed, algal-Lithocodium-Bacinella boundstone units between 107-112 m suggest lowstand time separation comparable to 4th-order DS 4b-6b in Wadi el Assy margin facies.

The Wadi el Assy section supports the hypothesis that the data suggest orbital forcing. However, the last rocks seen here are deeper-water facies that are biostratigraphically dated as decidedly younger than Jabal Madar section DS 6a. Thus, we must look elsewhere to evaluate the hypothesis that the orbital forcing hypothesis is in agreement with bathymetric evidence from the rest of the Oman Albian sections.

In average deeper-water section in the nearby Wadi Bani Kharus (Figures 2, 3 and 4) provides a link between Wadi el Assy and the rest of Oman’s Albian sections. Discontinuity surfaces 3a and 4a allow correlation from Wadi Bani Kharus to Wadi el Assy (Figure 3; Immenhauser and Scott, 2002). Wadi Bani Kharus contains MBS I-III, which are recognized elsewhere (Figure 3).

As indicated by Immenhauser and Scott (2002; their figure 2), Wadi Bani Kharus contains none of the low-amplitude (#b) events prevalent in Wadi el Assy and taken here to represent 4th-order lowstands. Indeed, Immenhauser and Scott (2002) correlate Wadi el Assy discontinuity surfaces 14-16b with (3rd-order lowstand) deposition of MBS I (Figure 3). Thus, in correlating sea-level calculation ‘A.2’ (Figure 11) to Wadi Bani Kharus (Figure 12) and Jabal Madar, we take the (#a) series of discontinuity surfaces to be 3rd-order cycle boundaries, sometimes expressed as single discontinuity surfaces (e.g. DS 1a, 2a, 3a, 4a, 5a and 10a) and sometimes including 4th or higher-order complexity (e.g. DS 6a, 7a and 8a, 9a; Figures 3 and 12).

Stratigraphic Modeling

Whereas the sea level PFM inputs orbital forcing time series and outputs a calculated sea-level time series, software program SimStrat inputs sea-level time series, combines it with driving subsidence, isostasy, compaction of underlying section and sedimentation rules to output a simulated stratigraphic section (Figure 10).

For this paper, the output is three depth-dependent sedimentary facies. For these rocks, facies can be thought of as: (1) grainstone to mud-lean packstone; (2) mud-rich packstone to wackestone; and (3) mudstone. Each facies has a sedimentation rate parameter and a water-depth range over which that rate applies. At each model time-step, water depth and isostasy/compaction determine accommodation space. If the sediment surface is above sea level, no sediment is deposited. If the sediment surface is below sea level, accommodation space is filled at the appropriate sedimentation rate until either available space is filled, or available time has run out. When a deeper-water facies fills to the top of its depth range, sediment type and rate switch to the next-shallower facies, which is deposited until either available space is filled or available time has run out.

Our stratigraphic modeling of an observed section begins with explicit recognition of four basic constraints. First, all model parameterizations must be geologically reasonable. Second, we note the measured thickness of the section to be modeled. Third, we set height of initial sediment surface to match water-depth estimates at and around the base of the section under consideration. Often, we are starting at a subaerial surface, but we also pay attention to facies below and above that surface. Fourth,
Figure 12: Model result depicting sea level and sediment surface as a function of time for the Wadi Bani Kharus section as per Table 2 SimStrat PFM parameterization. Correlation of model result to stratigraphic discontinuity surfaces (DS) 1a to 6a is indicated. The difference between sea level and sediment surface is presented as water depth in Figure 13b.

we give similar attention to the relation of the sediment surface to sea level at the top of the section under consideration. Thus, a constraint on all solutions to the forward problem may be written:

\[
\text{Sediment} = \text{Subsidence} \pm \text{Water Depth}
\]  \hspace{1cm} (1)

where,

1. **Sediment** is the sum of sedimentary thickness between base and top of section under consideration;
2. **Subsidence** is the net effect of driving subsidence, isostasy, compaction of underlying section;
3. **Water Depth** is any change in estimated water depth from base to top of the section.

Any acceptable forward solution must meet this basic constraint. The quality of the match between intervening model facies and thickness, and observed facies and thickness, determines the best of forward solutions under consideration.

**Table 2**

| SimStrat PFM Parameterization Wadi Bani Kharus Section |
|-------------------------------------------------------|
| **General Features**                                  |
| Driving Subsidence Rate:                             | 4 m/my   |
| Compensation for Isostasy/Compaction:               | 1.3 times water accommodation space available |
| Initial Sediment Surface (1a):                       | -50 meter below ice free world sea level |
| **Facies Input**                                     | **Facies Output** |
| Depth Range                                           | Thickness (DS 1a to 6a) (meters) |
| High Energy:                                         | 0 to 1 meter | 0.5 |
| Intermediate:                                        | 1 to 15 meter | 21.1 |
| Low Energy:                                          | >15 meter | 58.2 |
| Sedimentation Rate                                   | Sum Model > | 79.8 |
| High Energy:                                         | 10 m/my | Observed > | 80 |
| Intermediate:                                        | 5 m/my |           |
| Low Energy:                                          | 4 m/my |           |
Intermediate Water Depths: Wadi Bani Kharus Data/Model Comparison

Figure 12 presents calculated sea level (blue) versus time for the Wadi Bani Kharus section using the parameters in Table 2. The relation of the blue line to the red line depicts water depth (blue above red in Figure 12), or subaerial exposure (red above blue in Figure 12).

In order for the reader to appreciate our expectations of the model, we move up the section quickly comparing Figures 12 and 3, and commenting on model-dependent successes and shortcomings. Discontinuity surface 1a (Figure 3) comes in above the Shu’aiba Formation sedimentary rocks, which are indicative of on-average shallow-water conditions. We cannot model this. Consistent with previous work (Harris et al., 1984; Immenhauser et al., 1999; Sharland et al., 2001; Immenhauser et al., 2004), we suggest that the hiatus (several my) between the lower Aptian Shu’aiba and the Albian Nahr Umr formations is mainly the result of a plate-wide tectonic event in Arabia resulting in non-deposition and/or erosion of upper Aptian strata over much of Oman. Post-DS 1a, the section abruptly becomes low-energy facies. The model fit is better here. The high-energy conditions preceding DS 3a are adequately depicted in Figure 12. Likewise, DS 4a is modeled as topping MBS I and not indicating subaerial exposure.

However, at DS 5a, the target data indicate subaerial exposure; whereas the model does not quite reach subaerial exposure. This is a good example of our modeling strategy. We choose one set of parameters to apply over a long time interval (here, the entire section). If we set initial sediment surface so as to get subaerial exposure at DS 5a, DS 3a would be modeled as multiple subaerial surfaces. Rather than change parameterizations often to more nearly match the target data, we consider a certain amount of random variability and/or other processes to be geologically reasonable. Specifically, we acknowledge it is geologically unlikely that the Albian relative sea-level record of Oman reflects a 100% orbital-forcing pattern of regressions and transgressions without random variations in sedimentation rates, spasmodic tectonism, earthquakes etc. Thus, to attempt to model every last detail as the result of orbital forcing would, in our opinion, be superfluous.

Continuing up-section Wadi Bani Kharus, the comparison regards DS 6a/MBS II and DS 8a/MBS III target data signatures and model output is perhaps more tenuous. This model run is chosen to capture the former and thus is a mismatch to the latter. With these facies, it is impossible to model both signatures with a single parameterization. We are comfortable with both as 3rd-order lowstands with 4th-order complications, but abrupt changes in global climate (and thereby uncertainty regarding sea-level calculation ‘A.2’) may characterize this time interval. Recall that this is the time interval of the orbital forcing instability discussed above.

From the observational (stratigraphers) perspective, the Wadi Bani Kharus section is characteristically built by beds that have recessive, shaly and nodular carbonates at their base (see stratigraphic column for Wadi Bani Kharus in Figure 3). Towards the top of each bed, shaley limestones gradually change into clean(er) and resistant carbonates. We tentatively consider these shale-to-limestone successions as small-scale (4th and higher-order) ‘cycles’ and perhaps the intermediate water-depth equivalents of the shallow-water, small-scale cycles as recognized in the Wadi el Assy section (1b through 26b; Figure 3). In our interpretation, the shaly and nodular lower portion of these cycles represent the increased accommodation space during transgression (deep and muddy), and the clean(er) limestones of the upper portion stand for the decreasing accommodation space during regression. This pattern is in stratigraphic and spatial scale found in various orders of magnitudes up to the 100s-of-kilometer scale as reflected by the shaly intrashelf basinal sections contrasted by the clean-limestone platform top sections. Medium-scale (3rd-order) cycles in the Wadi Bani Kharus section are built by variable numbers of such shale-to-limestone cycles. The difficulty is the differentiation of 4th and higher-order events.

In Figure 13a, observational data is summarized as hydrodynamic level (Table 1) and plotted against height in measured section. Figure 13b shows modeled water depth (the difference between height of sea level and sediment surface in Figure 12) against modeled height in section, where the difference between DS 1a and 6a is set as a constraint (Equation 1) on the model solution. Thus, DS 1a and 6a may be regarded as input data, and height in section of other “#a” surfaces as model output. X-axis observation and model result are plotted so that subaerial exposure is to the left of the figure and deepwater is represented to the right margin of Figure 13.
In Figure 13, model spacing of DS 2a, 3a, and 4a closely approximates the observed spacing, and the maximum flooding intervals (MFI; light blue in Figure 13) are correlative. Consistent with observation, DS 1a, 2a and 3a are modeled as subaerial exposure. Discontinuity 4a is modeled as shoaling only as clear evidence for terrestrial subaerial exposure is lacking in Wadi Bani Kharus. Nevertheless, in the section west of Wadi Bani Kharus (Figures 1 and 3), geochemical and cement petrographic evidence points to at least short-lived exposure. Modeling suggests some 3rd-order cycle boundaries (e.g. meter 30, 50 and 69 in Figure 13, and 112, 108, 103 Ma (OFT) in Figure 12) involve less regression than others nearby. This is marginally supported by observation.

Modeling placement and amplitude of DS 5a poses two problems. First, model DS 5a comes close, but does not touch subaerial exposure. From the observational perspective, DS 5a shows good evidence
for subaerial exposure (Immenhauser et al., 2000b). Second, model sedimentation is a little less than observed DS 4a to 5a and a little more than observed DS 5a to 6a. As discussed above, the entire section DS 1a to DS 10a is modeled with a single parameterization. We consider this field data-model deviation as the result of random variation in sedimentation rates, local erosion, transport and re-deposition of sediment and topographic effects. These processes are not considered by the model and are expected to play some role. We thus do not consider this discrepancy a problem here, but we explore it further.

In cycle 1a to 2a (Figures 3 ‘Wadi Bani Kharus section’, and 13), a total of 12 shale-to-limestone beds is observed, but the thickness of these units varies between 0.1 and 1.2 m (black horizontal lines to the left of stratigraphic column indicate cycle tops). Judging from field observations and bed thickness, however, five shale-to-limestone successions stand out. This is consistent with the model (see Figure 12). The same accounts for cycle 2a to 3a (Figure 3, Wadi Bani Kharus section), but then with a seemingly persistent shallowing of facies. Here, five shale-to-limestone successions are recognized of which cycles 1 and 3 are built by six and three higher-order cycles, respectively. The model predicts 3 or 4 normal-amplitude 4th-order cycles followed by three low-amplitude cycles preceding DS 3a, a total of 6 to 7 cycles. This difference in cycle numbers (5 opposed by the predicted 6-7) is perhaps better understood when taking the higher-order cycles that build cycles 1 and 3 into consideration.

The section DS 3a-4a exhibits at least 10 cycles of variable thickness (Figure 3). Between meter 30 and 36 two firmgrounds (dotted lines in stratigraphic column) and intense bioturbation point to overall deeper conditions. Above meter 36, cycle count in Figures 3 and 12 is less straightforward. For the 3a to 4a interval, the model predicts eleven or twelve 4th-order cycles. The section DS 4a-5a exhibits 10 to 12 cycles, depending on what is considered 4th-versus higher order. For this interval, the model predicts thirteen 4th-order cycles (Figures 12 and 13). Thus, we accept this as a fair agreement between model and data.

The section DS 5a-6a exhibits 5 cycles to section meter 75, a little less than 2 m/cycle (Figures 3, 12 and 13). From 75 to 79 m, no cycles are indicated in Figure 13a, but perhaps two cycles maximum might be indicated by immediately preceding sedimentation rates; thus, a total of seven cycles maximum for this interval. The model predicts eleven 4th-order cycles for this interval. This is a discrepancy between model and data. Attempts to reconcile the problem in deeper-water sections proved inconclusive. Thus, we tentatively conclude the problem lies with the model.

Above DS 6a, we find a general correlation between model and data but the details are imperfect (Figures 12 and 13). From the observational perspective, the interval between DS 6a and 7a is characterized by what we consider the deep-water facies beneath the storm wave base, as confirmed by the biota

### Table 3

| SimStrat PFM Parameterization Wadi el Assyir margin section |
|-----------------------------------------------------------|
| **General Features**                                      |
| Driving Subsidence Rate: 7 m/my                           |
| Compensation for Isostasy/Compaction: 2.2 times water accommodation space available |
| Initial Sediment Surface (3a): -38 meter below ice free world sea level |

| **Facies Input** | **Depth Range** | **Thickness (DS 3a to 13b)** (meters) |
|------------------|-----------------|--------------------------------------|
| High Energy:     | 0 to 10 meter   | 50.3                                 |
| Intermediate:    | 10 to 15 meter  | 8.3                                  |
| Low Energy:      | >15 meter       | 1                                    |

| **Sedimentation Rate** | **Depth Range** | **Thickness (DS 3a to 13b)** (meters) |
|------------------------|-----------------|--------------------------------------|
| High Energy:           | 30 m/my         | 59.6                                 |
| Intermediate:          | 15 m/my         | Observed > 59                        |
| Low Energy:            | 30 m/my         |                                       |
including ammonites. We tentatively correlate DS 8a/9a observed with model meter 88 (Figure 13).
While we consider the close spacing of two surfaces to be a common signature of 3rd-order lowstands,
about 10 m of evaporative sea-level drawdown or other process seems required. Surface 10a poses a
similar problem. Correlated to meter 97 in model section (~96 Ma (OFT)), approximately 15 m alternative
process seems required. Correlated to meter 108 in model section (~93 Ma (OFT)), and we are putting
too much reliance on the segment of the model that was found to contain an instability.

In Wadi Bani Kharus, the interval above DS 8a is characterized by an overall decreasing accommodation
space reflected by the well-washed massive limestones (locally cross-bedded) that characterize the
transition from the Albian Nahr Umr Formation to the overlying Cenomanian Natih Formation
(Immenhauser et al., 1999). The onset of the Natih Formation reflects the end of the Nahr Umr intrashelf
basinal sedimentation, and an overall shallower depositional environment (van Buchem et al., 1996). For
orbital forcing to accommodate these observations would require (a) correlating DS 8a with something
above 94 Ma (OFT) and (b) significant change in model parameterization commencing above DS 6a
(Figure 12). Consistent with our strategy (above), we do not undertake such steps at this time.

Summarizing our modeling of the Wadi Bani Kharus section, a single parameterization of the model
yielded acceptable model/data comparison for the first ten million years Ma (OFT) of the model run.
That is a long time to expect the Earth to remain so much the same. We note the 4th-order cycle count for
the 3rd-order interval, DS 5a to 6a, as a discrepancy to be further evaluated (but unresolved) below.

Shallow Water Depths: Wadi el Assyi Data/Model Comparison

Parameterization for our modeling of Wadi el Assyi margin section is provided in Table 3. This
parameterization was applied from DS 3a to the top of the margin section. The output satisfies equation
1 for the interval DS 3a to 13b. Parameterization for our modeling of Wadi el Assyi lagoon section is
provided in Table 4. This parameterization was applied from DS 4a to the top of the section. The output
satisfies equation 1 for the interval DS 4a to 26b.

From the observational (stratigraphers) perspective, the Wadi el Assyi section does not resolve clear
differences in accommodation space, as much of these sedimentary rocks were deposited above the
effective wave base (Figures 3 and 4). Furthermore, as the Al Hassanat Platform prograded in mid-
Albian times, the one-dimensional section measured in Wadi el Assyi records the shift from the Al
Figure 15: Wadi el Assy platform margin stratigraphic section (left) compared to modeled water depth plotted as modeled height in section (right). A single parameterization (Table 3) is applied from DS 3a through DS 13b. From DS 3a to DS 7b, four 4th-order cycles of roughly equal thickness are observed and four 4th-order cycles of roughly equal thickness are modeled, though thickness runs consistently greater than observed. Note cycle 10b thin, 11b thick, 12b thin observed and thin/thick/thin modeled, though thickness again runs consistently greater than observed. See text for discussion of cycles 7b and 8b.
Hassanat Platform margin to the platform lagoon (Immenhauser et al., 2001). In the section, this results in an apparent net increase of water depth, as the platform margin stood higher than the carbonate seafloor of the lagoon. Obviously, this ‘deepening’ is not related to fluctuations in sea level but is a topographic feature. As shown in Figure 3 (Wadi el Assyi section), the margin (basically the interval from DS 2 and 3a to 4a) is predominantly built by rudist-coral lithosomes. Above surface 4a, the facies records the platform lagoon and small cycles built by low-energy lagoonal deposits grading into high-energy rudist facies might represent 4th- and higher-order, sea-level cycles. The interval between DS 13b and 4a is transitional between margin and lagoon.

Given the lateral limitations of these outcrops (Immenhauser et al., 2001), we cannot easily distinguish between autocyclic and allocyclic shoaling. We thus limit our interpretation to those cycles that are bound by exposure surfaces, a clear evidence of relative sea-level fall as marine carbonates do not build above sea level (Immenhauser and Scott, 2002). In Figure 3, DS 17b through 22b and 23b through 26b cap such shoaling cycles. Judging from the variable thickness of these cycles, the duration of the mid-Albian, and based on the overall shallow setting, we assume that cycles of different orders are recorded.

Figure 5c explains the evolution of discontinuity surfaces, the terminology and margin/slope relations in Wadi el Assyi. Figure 14 presents a plot for the Wadi el Assyi margin section of model input/output of similar construct to Figure 12 (Wadi Bani Kharus). Figure 15 compares the Wadi el Assyi margin measured section with model-generated water depth estimates and height in section. The margin section from DS 3a to 13b is modeled with a single parameterization (Table 3). Thus, DS 3a and 13b may be regarded as input data (i.e. satisfying equation 1) and height in section of other “#b” surfaces as model output.

As water-depth estimates are deemed unreliable for shallow, high-energy facies such as the Wadi el Assyi margin, the thickness between discontinuity surfaces is about all we have to work with. We characterize the observed section DS 3a to 8b as five 4th-order cycles of about equal thickness (Figure 15). We model DS 3a to 7b as four 4th-order cycles of about equal thickness. Note in Figure 14 that cycle 7b is a 4th-order cycle that is poorly represented in model output because of subaerial exposure. Assuming factors such as for instance earthquakes to play a role, this interval could easily look like the other modeled cycles 3a to 7b.

Higher up-section, observed section DS 10b to 13b consists of a thick cycle (11b) with thinner cycles (10b and 12b), above and below. In Figure 13 that 4th-order cycle 11b has ‘broad shoulders’ compared to any other cycle in this model run. We comment further in discussion.

### Table 4

| SimStrat PFM Parameterization Wadi el Assyi lagoon section |
|-----------------------------------------------------------|
| **General Features**                                      |
| Driving Subsidence Rate:                                  | 11 m/my |
| Compensation for Isostasy/Compaction:                    | 2.1 times water accommodation space available |
| Initial Sediment Surface (4a):                           | -40 meter below ice free world sea level |
| **Facies Input**                                          |
| **Facies Output**                                         |
| **Depth Range**                                           |
| High Energy: 0 to 1 meter                                 | 2.3 |
| Intermediate: 1 to 15 meter                               | 59.3 |
| Low Energy: >15 meter                                     | 19.3 |
| **Sedimentation Rate**                                    |
| High Energy: 50 m/my                                      | Observed > 80 |
| Intermediate: 20.9 m/my                                   | 80 |
| Low Energy: 10.5 m/my                                     | 80 |

Downloaded from http://pubs.geoscienceworld.org/georarabia/article-pdf/9/3/11/5441861/immenhauser.pdf by guest
The above discussion leaves us with cycle 8b (the stratigraphic interval between DS 8b and 9b in Figure 3) unexplained. Observation has 8b as the thickest cycle in the margin sequence; modeling has it as one of the thinnest. Either the modeling concept of a “signature” to 3rd and 4th-order cycles is not working out, or one (or more) discontinuity surface(s) between DS 8b and 9b has (have) escaped our attention in the field. This is not unlikely given the subtle nature of these features.

Modeling of the Wadi el Assyi lagoon section (Table 4) came out very similar to Figure 13, the sediment surface running about 15 m shallower than in the Wadi Bani Kharus section. Discontinuity surfaces 17b to 22b are correlated with 3rd-order low stand at ~112 Ma (OFT) and DS 23b to 26b with 3rd-order

| Inferred hydrodynamic level | Outcrop Jabal Madar | Exploration well | Sea level (meter below ice free) |
|-----------------------------|---------------------|------------------|---------------------------------|
| high intermittent low        |                      |                  |                                 |
| high intermittent low        |                      |                  |                                 |

Figure 16: Comparison of rock record of marker bed successions (MBS) I and II from Jabal Madar, with total gamma ray from nearby Jabal Madar exploration well and the modeled sea-level curve. An interpretation of matching peaks and lows is indicated using numerals 1 to 9 for MBS II (16a) and 1 to 13 for MBS I (16b). The Jabal Madar section reflects a shaley, below-storm wave base, depositional setting (brown color) punctuated by an upward shoaling limestone succession (MBS, indicated in yellow) capped by discontinuity surfaces (DS) 4a and 6a respectively showing evidence for terrestrial subaerial exposure. Each carbonate bed (green/blue or gray respectively) indicates a relative fall; the alternating shaley interbeds a relative rise in sea level (4 to 8 for MBS II and 6 to 9 for MBS I). This pattern perhaps reflects the superposition of higher-order sea-level fluctuations on an overall regressive trend. A very comparable pattern is present in the total gamma ray and in the modeled sea-level curve. Note variant gamma ray and sea-level curve beneath and above of MBS’s. This refers to a deep basinal interval in the rock record. It is conceivable that this variant pattern reflects differences in organic content, which in turn are related to higher-order sea-level fluctuations as indicated in the model curve. Refer to Figure 3 for a more detailed key to colors and symbols.
low stand at ~105 Ma (OFT). The modeled section ends with water getting deeper at ~102 Ma (OFT), just short of encountering MBS II (Figure 15).

**Deep Intrashelf Basin: Jabal Madar Data/Model Comparison**

Figure 16 compares portions of the Jabal Madar section (Figure 3) with total gamma-ray data from the nearby Jabal Madar exploration well and to our sea-level model output. From the stratigraphers perspective, the bathymetric information obtained in the Jabal Madar section is limited as much of the section is built by barren shales that are indicative for water depths beneath the effective storm wave base. As shown in Figure 3, the shales are interrupted by three calcareous marker bed successions topped by subaerial exposure surfaces (DS 4a, 6a, and 8a). We use these calcareous intervals as reference in Figure 16 as they provide information about changes in the hydrodynamic level. The characteristic feature of both MBS I and MBS II is that they are built by four and two thick calcareous units, respectively, that alternate with shaley interbeds (Figures 3 and 15). The boundaries between limestone beds and shales are sharp but this might be the result of carbonate diagenesis. The lithofacies and biota of these limestone beds is indicative for a stepwise shoaling and a stepwise increase in hydrodynamic energy (Immenhauser et al., 1999; Figure 16).

The total gamma-ray signatures from these intervals could reflect subtle changes in clay-carbonate and organic matter content. We have, however, no access to spectral gamma ray data that would allow for a differentiation between clay and organic matter content. The gamma-ray signature shows a remarkable variability in the apparently monotonous shale intervals beneath marker bed successions. These fluctuations may reflect environmental changes, perhaps driven by climate and thus eventually sea-level fluctuations.

Comparing the lithofacies with the gamma-ray signature and the sea-level file reveals a strikingly similar pattern. We show a tentative interpretation of matching highs/lows in Figure 15 and indicate correlative peaks with numerals.

Discontinuity 4a is abruptly overlain by shale, whereas the change from limestone to shale is more gradual above DS 6a. In this respect, we see a good agreement between model and data (gamma ray and outcrop).

Discontinuity 5a is not expressed at Jabal Madar and not exposed in the Wadi Mu‘aydim section (Figure 3). In contrast, the Southern Huqf section in Figure 3, records a number of calcareous beds between DS 4a and 6a that might represent the intrashelf basinal equivalent of a sea-level fall related to the DS 5a event. Using the Jabal Madar exploration well gamma ray section and assuming that ‘saw-tooth’ shaped gamma ray ‘cycles’ reflect subtle differences in carbonate and/or organic matter content driven by sea-level fluctuations, a very tentative assessment of cycle numbers is possible. The gamma-ray section displays perhaps seven or eight 4th-and higher-order cycles for the interval DS 4a to 5a. For the interval DS 5a to 6a, perhaps 4 to 5 cycles are visible. The cycle count for DS 4a to 5a being less than observations at the Wadi Bani Kharus section renders this observation inconclusive in regards to the DS 5a to 6a interval. The DS 5a to 6a problem may reside with the model or it may reside with the sediments at localities studied thus far being imperfect recording systems.

**CONCLUSIONS**

(1) Using the frequency and amplitude of relative 3rd, 4th and higher-order Albian sea-level events in Oman as a ‘Rosetta Stone’, we have compared these characteristics with the frequency and amplitude of all published mechanisms that cause sea-level change in a similar order-of-magnitude. Stress-related deformation of margins result in sea-level changes with amplitudes similar to the Oman sea-level cycles but has an insufficient spatial component (200-250 km as opposed by the 500 km of the Oman transect). Similarly, ridge volume related changes generally result in sea-level oscillations that had lower amplitudes and lower frequencies than the Oman sea-level curve. From all processes considered, glacio-eustasy (or unknown processes we do not understand) agrees best with the frequency and amplitudes of the Oman cycles. This is a controversial conclusion and suggests re-evaluation of the Cretaceous Period as one of global, continuous climatic warmth.
2) We have investigated orbital forcing of glacio-eustasy as an explanation of the data. We find numerous observational patterns of orbital forcing, but do not achieve an exact match, particularly for the interval above discontinuity surface 5a.

3) We definitely do not propose orbital glacio-eustasy as the sole mechanisms that influenced the Albian sea level in Oman. Variable uplift and subsidence on local to regional scales, variable sedimentation rates, sediment erosion, lateral sediment transport and re-deposition, and other local mechanisms, shall remain a concern for the observational stratigrapher. Some of these concepts can be built into stratigraphic modeling with ease.

4) Patterns of orbital forcing observed here, and to be expected in similar studies include: (a) overwhelming 4th and higher-order cycles in shallow-water sections, with bundling of 4th-order into 3rd-order becoming apparent in somewhat deeper-water sections. (b) Bundling of five to seven 4th-order cycles to form $2.4 \pm 0.4$ my, or eleven to thirteen 4th-order, to form $4.8 \pm 0.4$ my cycles. This is expected to constitute a veritable “Morse Code” of short (3rd-order) and long (two 3rd-order cycles) that should be robust. (c) Bursts of 4th- and higher-order cycles at 3rd-order sequence boundaries, which may appear as marker bed sequences and/or subaerial exposure surfaces.

5) Whereas biostratigraphy focuses on the history of life, which may or may not reflect sequence boundaries, orbital forcing focuses exclusively on events that should be reflected in sequence stratigraphy. Usage of ‘orbital forcing time’ as relative time units helped these authors past some impasses, and should be given serious consideration.

6) The outcome of this data/model comparison of the Oman Albian sea-level curve suggests that further collaboration between stratigraphers and modelers is clearly promising. The main aspects of further work should include an improved convergence between biostratigraphic/radiometric time scales with orbital forcing time, and a better separation of 4th and higher-order cycles both in the field and in the model output. Obviously, the focus must be on sequence boundaries rather than on classical European stage names and on biological events in taxa far removed from the stage name type localities.

ACKNOWLEDGEMENTS

We acknowledge Petroleum Development Oman and the Ministry of Oil and Gas of Oman for access to gamma ray data from the Jabal Madar exploration well. We appreciate all constructive comments from two anonymous GeoArabia reviewers and regret that we disagree in some fundamental aspects. The authors are indebted to GeoArabia Editor Moujahed I. Al-Husseini for the editorial handling of this manuscript. The final design and drafting of graphics by Gulf PetroLink is acknowledged.

REFERENCES

Alsharhan, A.S. 1995. Facies variation, diagenesis, and exploration potential of the Cretaceous rudist-bearing carbonates of the Arabian Gulf. American Association of Petroleum Geologists, v. 79, p. 531-550.

Berger, A. and M.F. Loutre 1991. Insolation values for the climate of the last 10 million years. Quaternary Science Reviews, v. 10, p. 297-317.

Berger, A., M.F. Loutre and J. Laskar 1992. Stability of the astronomical frequencies over the Earth’s history for paleoclimate studies. Science, v. 255, p. 560-566.

Bryan, K. and M.D. Cox 1972. An approximate equation of state for numerical models of ocean circulation. Journal of Physical Oceanography, v. 2, p. 510-517.

Burton, R., C.S.C. Kendall and I. Lerche 1987. Out of our depth: on the impossibility of fathoming eustasy from the stratigraphic record. Earth-Science Reviews, v. 24, p. 237-277.

Carney, J.L. and R.W. Pierce 1995. Graphic correlation and composite standard databases as tools for the exploration biostratigrapher. In, K.O. Mann and H.R. Lane (Eds.), Graphic Correlation. SEPM Special Publication v. 53, p. 23-43.

Clark, J.A., W.E. Farrell and W.R. Peltier 1978. Global changes in postglacial sea-level: a numerical calculation. Quaternary Research, v. 9, p. 265-287.
Cloeingh, S., H. McQueen and K. Lambeck 1985. On a tectonic mechanism for regional sea-level variations. Earth and Planetary Science Letters, v. 75, p. 157-166.
Cloeingh, S. 1986. Intraplate stresses: a new tectonic mechanism for fluctuations of relative sea level. Geology, v. 14, p. 617-620.
Davies, G.F. 1984. Lagging mantle convection, the geoid and mantle structure. Earth and Planetary Science Letters, v. 69, p. 187-194.
Dewey, J.F. and W.C. Pitman 1998. Sea-level changes: mechanisms, magnitudes and rates. In, J.L. Pindell and C. Drake (Eds.), Paleoceanographic Evolution and Non-glacial Eustasy. SEPM Special Publication v. 58, p. 2-16.
Douglas, J.G. and G.E. Williams 1982. Southern polar forests: the early Cretaceous floras of Victoria and their paleoclimatic significance. Palaeogeography, Palaeoclimatology, Palaeoecology, v. 94, p. 261-282.
Dubreuilh, J., J.-P. Platel, J. Le Métour, R. Wyns, F. Béchennec and A. Bertiaux 1992. Geological map of Khaluf with explanatory notes, sheet NF 40-15. Ministry of Petroleum and Minerals, Directorate General of Minerals, Muscat, Oman.
Eyles, N. 1993. Earth's glacial record and its tectonic setting. Earth-Science Reviews, v. 35, p. 1-248.
Fastook, J.L. and M. Prentice 1994. A finite-element model of Antarctica - sensitivity test for meteorological mass-balance relationship. Journal of Glaciology, v. 40, p. 167-175.
Frakes, L.A. and J.E. Francis 1988. A guide to Phanerozoic cold polar climates from high-latitude ice-rafting in the Cretaceous. Nature, v. 333, p. 547-549.
Frakes, L.A., N.F. Alley and M. Dynek 1995. Early Cretaceous ice rafting and climate zonation in Australia. International Geological Review, v. 37, p. 567-583.
Frakes, L.A. 1999. Estimating the global thermal state from Cretaceous sea surface and continental temperature data. In, E. Barrera and C.C. Johnson (Eds.), Evolution of the Cretaceous Ocean-Climate System. Geological Society of America, Special Paper v. 332, p. 49-57.
Francis, J.E. and L.A. Frakes 1993. Cretaceous climates. In, V.P. Wright (Ed.), Sedimentology Review. Blackwell, v. 1, p. 17-30.
Glennie, K.W., M.G.A. Boeuf, M.W. Hughes Clarke, M. Moody-Stuart, W.F.H. Pilaar and B.M. Reinhardt 1974. Geology of the Oman Mountains. Shell Research BV, The Hague, 423 p.
Gradstein, F.M., F.P. Agterberg, J.G. Ogg, J. Hardenbol, P. van Veen, J. Thierry and Z. Huang 1994. A Mesozoic time scale. Journal of Geophysical Research, v. 99(B12), p. 24,051-24,074.
Gradstein, F.M., F.P. Agterberg, J.G. Ogg, J. Hardenbol, P. van Veen, J. Thierry and Z. Huang 1995. A Triassic, Jurassic and Cretaceous time scale. In, W.A. Berggren, D.V. Kent, M.-P. Aubry and J. Hardenbol (Eds.), Geochronology, Time Scales and Global Stratigraphic Correlation. SEPM Special Publication v. 54, p. 95-126.
Hallam, A. 1981. A revised sea level curve for the Early Jurassic. Quarterly Journal of the Geological Society of London, v. 138, p. 735-743.
Hallam, A. 1992. Phanerozoic Sea-Level Changes. Columbia University Press, New York, 224 p.
Haq, B.U. and F.W.B. Van Eysinga 1987. Geological Time Table (4th edition, Wall Chart). Elsevier, Amsterdam.
Haq, B.U., J. Hardenbol and P.R. Vail 1987. Chronology of the fluctuating sea levels since the Triassic. Science, v. 235, p. 1156-1167.
Hardenbol, J., J. Thierry, M.B. Farley, T.Jacquin, P.-C. DeGraciansky and P.R. Vail 1998. Mesozoic-Cenozoic sequence chronostratigraphic framework. In, P.-C. DeGraciansky, J. Hardenbol, T. Jacquin, P.R. Vail and M.B. Farley (Eds.), Sequence Stratigraphy of European Basins. Society for Sedimentary Geology, SEPM Special Publication v. 60, p. 3-13.
Harland, W.B., R.L. Armstrong, A.V. Cox, L.E. Craig, A.G. Smith and D.G. Smith 1990. A Geologic Time Scale. Cambridge University Press, New York, 265 p.
Harris, P.M. and S.H. Frost 1984. Middle Cretaceous Carbonate reservoirs, Fahud field and northeastern Oman. American Association of Petroleum Geologists Bulletin, v. 68, p. 649-658.
Harris, P.M., S.H. Frost, G.A. Seiglie and N. Schneidermann 1984. Regional unconformities and depositional cycles, Cretaceous of the Arabian Peninsula. In, J.S. Schlee (Ed.), Interregional Unconformities and Hydrocarbon Accumulation. American Association of Petroleum Geologists Memoir, v. 36, p. 67-80.
Harrison, C.G.A. 1990. Long-term eustasy and epeirogeny in continents. In, N.R. Council (Ed.), Sea-level Change. National Academy Press, Washington, DC, Studies in Geophysics, p. 141-160.
Harrison, C.G.A., G.W. Brass, B. Saltzman, J. Sloan II, J. Southam, and J.M. Whitman 1981. Sea level
variations, global sedimentation rates, and the hypsographic curve. Earth and Planetary Science Letters, v. 54, p. 1-16.
Heller, P.F., D.L. Anderson and C.L. Anjevine 1996. Is the middle Cretaceous pulse of rapid sea-floor spreading real or necessary? Geology, v. 24, p. 491-493.
Hoffman, P.F. and D.P. Schrag 2000. Snowball Earth. Scientific American, v. 282, p. 68-75.
Hughes Clarke, M.W. 1988. Stratigraphy and rock unit nomenclature in the oil-producing area of Interior Oman. Journal of Petroleum Geology, v. 11, p. 5-60.
Immenhauser, A., G. Schreurs, T. Peters, A. Matter, M. Hauser and P. Dumitrica 1998. Stratigraphy, sedimentology and depositional environments of the Permian to uppermost Cretaceous Batin Group, eastern-Oman. Eclogae geologicae Helvetiae, v. 91, p. 217-236.
Immenhauser, A. and R.W. Scott 1999. Global correlation of middle Cretaceous sea-level events. Geology, v. 27, p. 551-554.
Immenhauser, A., W. Schlager, S.J. Burns, R.W. Scott, T. Geel, J. Lehmann, S. van der Gaast and L.J.A. Bolder-Schrijver 1999. Late Aptian to Late Albian sea-level fluctuations constrained by geochemical and biological evidence (Nahr Umr Fm, Oman). Journal of Sedimentary Research, v. 69, p. 434-446.
Immenhauser, A., W. Schlager, S.J. Burns, R.W. Scott, T. Geel, J. Lehmann, S. van der Gaast and L.J.A. Bolder-Schrijver 2000a. Origin and correlation of disconformity surfaces and marker beds, Nahr Umr Formation, northern Oman. In, A.S. Alsharhan and R.W. Scott (Eds.), Middle East Models of Jurassic/Cretaceous Carbonate Systems. SEPM Special Publication v. 69, p. 209-225.
Immenhauser, A., A. Creusen, M. Esteban and H.B. Vonhof 2000b. Recognition and interpretation of polygenic discontinuity surfaces in the Middle Cretaceous Shu’aiba, Nahr Umr, and Natih formations of northern Oman. GeoArabia, v. 5, no. 2, p. 299-322.
Immenhauser, A., B. van der Kooij, A. van Vliet, W. Schlicher and R.W. Scott 2001. An ocean facing Aptian-Albian carbonate margin, Oman. Sedimentology, v. 48, p. 1187-1207.
Immenhauser, A. and R.W. Scott 2002. An estimate of Albian sea-level amplitudes and its implications for the duration of stratigraphic hiatuses. Sedimentary Geology, v. 152, p. 19-28.
Immenhauser, A., H. Hillgärtner, U. Sattler, G. Bertotti, P. Schoepfer, P. Homewood, V. Vahrenkamp, T. Steuber, J.-P. Masse, H.H.J. Droste, J. van Koppen, B. van der Kooij, van E.C. Bentum, K. Verver, E. Hoogenderduijn-Strating, W. Swinkels, J. Peters, I. Immenhauser-Potthast and S.A.J. Al-Maskery 2004. Barremian-lower Aptian Qishn Formation, Haushi-Huqf area, Oman: a new outcrop analogue for the Kharaib/Shu’aiba reservoirs. GeoArabia (in press February 2004).
Intergovernmental Panel on Climate Change, W.G.I. Climate Change 2001: The Scientific Basis, 873 p. United Nations Environment Programme (UNEP), Hertfordshire, UK.
Kaiho, K. and S. Saito 1994. Oceanic crust production and climate during the last 100 Myr. Terra Nova, v. 6, p. 376-384.
Komintz, M.A. 1984. Oceanic ridge volumes and sea-level change - an error analysis. In, J.S. Schlee (Ed.), Interregional Unconformities and Hydrocarbon Accumulation. American Association of Petroleum Geologists Memoir, v. 36, p. 109-127.
Lambeck, K. 1988. Geophysical geodesy: the slow deformations of the earth? Oxford Scientific Publications, Oxford.
Larson, R.L. and C. Kincaid 1996. Onset of mid-Cretaceous volcanism by elevation of the 670 kilometer thermal boundary layer. Geology, v. 24, p. 551-554.
Larson, R.L. 1991. Geological consequences of superplumes. Geology, v. 19, p. 963-966.
Laskar, J. 1990. The chaotic motion of the Solar System: a numerical estimate of the size of the chaotic zones. Icarus, v. 88, p. 266-291.
Laskar, J. 1999. The limits of Earth orbital calculations for geological time-scale use. Philosophical Transactions of the Royal Society of London, v. 357, p. 1735-1760.
Masse, J.-P., J. Borgomano and S. Al-Maskiry 1997. Stratigraphy and tectonostratigraphic evolution of an upper Aptian-Albian carbonate margin: the northeastern Jabal Akhdar (Sultanate of Oman). Sedimentary Geology, v. 113, p. 269-280.
Matthews, R.K. and C. Frohlich 1998. Forward modeling of sequence stratigraphy and diagenesis: application to rapid, cost-effective carbonate reservoir characterization. GeoArabia, v. 3, no. 3, p. 359-384.
Matthews, R.K. and C. Frohlich 2002. Maximum flooding surfaces and sequence boundaries: comparisons between observations and orbital forcing in the Cretaceous and Jurassic (65-190 Ma). GeoArabia, v. 7, no. 3, p. 503-538.
Mattner, J. and M.I. Al-Husseini 2002. Essay: applied cyclo-stratigraphy for the Middle East E&P industry. GeoArabia v. 7, no. 4, p. 734-744.

McNutt, M.K., E.L. Winterer, W.W. Sager, J.H. Natland and G. Ito 1990. The Darwin Rise: a Cretaceous superswell? Geophysical Research Letters, v. 17, p. 1101-1104.

Mörner, N.-A. 1976. Eustasy and geoid changes. The Journal of Geology, v. 84, p. 123-151.

Mörner, N.-A. 1980. Relative sea-level, tectono-eustasy, geoidal-eustasy and geodynamics during the Cretaceous. Cretaceous Research, v. 1, p. 329-340.

Mörner, N.-A. 1981. Revolution in Cretaceous sea-level analysis. Geology, v. 9, p. 344-346.

Mörner, N.-A. 1987. Models of global sea-level changes. In, M.J. Tooley and I. Shennan (Eds.), Sea Level Changes. Blackwell, Oxford, p. 332-355.

Murris, R.J. 1980. Middle East: stratigraphic evolution and oil habitat. American Association of Petroleum Geologists Bulletin, v. 64, p. 597-618.

Pratt, B.R. and J.D. Smewing 1993. Early Cretaceous platform-margin configuration and evolution in the Central Oman Mountains, Arabian Peninsula. American Association of Petroleum Geologists Bulletin, v. 77, p. 225-244.

Prentice, M.L. and R.K. Matthews 1991. Tertiary ice sheets dynamics: the Snow Gun hypothesis. Journal of Geophysical Research, v. 96, p. 6811-6827.

Price, G.D. 1999. The evidence and implication of polar ice during the Mesozoic. Earth-Science Reviews, v. 48, p.183-210.

Pugh, D.T. 1993. Improving sea-level data. In, R.A. Warrick (Ed.), Climate and Sea Level Change: Observations, Projections and Implications. Cambridge University Press, Cambridge, p. 57-71.

Quinn, T.M., S. Tremaine and M. Duncan 1991. A three million year integration of the Earth’s orbit. Astronomical Journal, v. 101, p. 2287-2305.

Rabu, D. 1987. Géologie de l’Autochtone des montages d’Oman: la fenetre du Jabal Akhdar. La Semelle Métamorphique de la Nappe Ophiolitique de Semail dans les Parties Orientale et Centrale des Montagnes d’Oman: une Revue. Doctoral thesis, Pierre and Marie Curie University, Paris 6. French Bureau de Recherches Géologiques et Minières Document no. 130, 582 p.

Ravaut, P., A. Alyahyaey, R. Bayer and A. Lesquer 1993. Isostatic response of the Arabian Platform to Ophiolitic Loading in Oman. Comptes Rendues de L’Academie des Sciences Serie II, v. 317, p. 463-470.

Ries, A.C. and R.M. Shackleton 1990. Structures in the Huqf-Haushi Uplift, east central Oman. In, A.H.F. Robertson, M.P. Searle and A.C. Ries (Eds.), The Geology and Tectonics of the Oman Region. Geological Society of London, Special Publication v. 49, p. 653-664.

Scotese, C.R. 2001. Atlas of Earth History, Volume 1, Paleogeography, PALEOMAP Project, Arlington, Texas. v. 1, 52 p.

Scott, R.W. 1990. Chronostratigraphy of the Cretaceous carbonate shelf, southeastern Arabia. In, A.H.F. Robertson, M.P. Searle and A.C. Ries (Eds.), The Geology and Tectonics of the Oman Region. Geological Society of London, Special Publication v. 49, p. 89-108.

Sellwood, B.W., G.D. Price and P.J. Valdes 1994. Cooler estimates of Cretaceous temperatures. Nature, v. 370, p. 453-455.

Sharland, P.R., R. Archer, D.M. Casey, R.B. Davies, S.H. Hall, A.P. Heward, A.D. Horbury and M.D. Simmons 2001. Arabian Plate Sequence Stratigraphy. Gulf PetroleumLink, Manama, Bahrain, 371 p.

Shaw, A.B. 1964. Time in Stratigraphy. McGraw-Hill, New York. 365 p.

Siddall, M., E.J. Rohling, A. Almogi-Labin, C. Hemleben, D. Meischner, I. Schmelzer and D.A. Smeed 2003. Sea-level fluctuations during the last glacial cycle. Nature, v. 423, p. 853-858.

Simmons, M.D. and M.B. Hart 1987. The biostratigraphy and microfacies of the Early to mid-Cretaceous carbonates of Wadi Mi’aidin, Central Oman Mountains. In, M.B. Hart (Ed.), Micropaleontology of Carbonate Environments. Ellis Horwood, Chichester, UK. p. 176-207.

Stoll, H.M. and D.P. Schrag 1996. Evidence for glacial control of rapid sea-level changes in the Early Cretaceous. Science, v. 2, p. 1771-1774.

Sykes, T.J.S. and R.B. Kidd 1994. Volcanogenic sediment distribution in the Indian Ocean through the Cretaceous and Cenozoic, and their paleoenvironmental implications. Marine Geology, v. 116, p. 267-291.

Tamisiea, M.E., J.X. Mitrovica, G.A. Milne and J.L. Davis 2001. Global geoid and sea-level changes due to present-day ice mass fluctuations. Journal of Geophysical Research, v. 106, p. 30,849-30,863.

Tooley, M.J. 1993. Long term changes in eustatic sea level. In, R.A. Warrick, E.M. Barrow and T.M.L. Wigley (Eds.), Climate and Sea Level Change. Cambridge University Press, Cambridge. p. 81-110.
Immenhauser and Matthews

Vail, P.R., R.M.J. Mitchum and S. Thompson 1977. Seismic stratigraphy and global changes of sea-level. Part 3 Relative changes of sea-level from coastal onlap. In, C. Payton (Ed.), Seismic Stratigraphy. American Association of Petroleum Geologists Memoir, v. 26, p. 63-81.

van Buchem, F.S.P., P. Razar, P.W. Homewood, J.M. Philip, G.P. Eberli, J.-P. Platel, J. Roger, R. Eschard, G.M.J. Desabliaux, U.T. Boisseauf, J.-P. Leduc, R. Loubardette and S. Cataloube 1996. High resolution Sequence Stratigraphy of the Natih Formation (Cenomanian/Turonian) in Northern Oman: distribution of source rocks and reservoir facies. GeoArabia, v. 1, no. 1, p. 65-91.

van Buchem, F., B. Pittet, H. Hillürtner, J. Grötsch, A.I. Al-Mansouri, I.M. Billing, H.J. Droste and W.H. Oterdoom 2002. High-resolution sequence stratigraphic architecture of Barremian/Aptian Carbonate Systems in northern Oman and the United Arab Emirates (Kharaib and Shu’aiba formations). GeoArabia, v. 7, no. 3, p. 461-500.

Warrick, R.A. 1993. Climate and sea level change: a synthesis. In, R.A. Warrick, E.M. Barrow and T.M.L. Wigley (Eds.), Climate and Sea Level Change: Observations, Projections and Implications. Press Syndicate of the University of Cambridge, Cambridge. p. 3-24.

Weissert, H. and A. Lini 1991. Ice age interludes during the time of Cretaceous greenhouse climate. In, D.W. Müller, J.A. McKenzie and H. Weissert (Eds.), Controversies in Modern Geology - Evolution of Geologic Theories in Sedimentology, Earth History and Tectonics. Academic Press. p. 173-190.

Wigley, T.M.L. and S.C.B. Raper 1993. Future change in global mean temperature and sea level. In, R.A. Warrick, E.M. Barrow and T.M.L. Wigley (Eds.), Climate and Sea Level Change. Cambridge University Press, Cambridge. p. 111-133.

ABOUT THE AUTHORS

Adrian Immenhauser is an Assistant Professor in the Department of Earth and Life Sciences of the Vrije Universiteit, Amsterdam, The Netherlands. He received his PhD from the University of Berne, Switzerland. His interest in the Geology of Oman started in 1992 when he became involved in an interdisciplinary study of the Masirah Island Ophiolites. His present regional research focus is on Cretaceous carbonate systems of the Middle East and Greece and Upper Carboniferous platforms in Europe. A long-term goal is the compilation of a consistent platform-wide synthesis of the mid-Cretaceous units in eastern Arabia. Two recent lines of research focus on the: (1) relation between facies, diagenesis and fracturing of carbonate reservoir facies and (2) the impact, recognition and interpretation of hydrothermal karsting in subsurface settings.

Robley (Rob) K. Matthews is Professor of Geological Sciences at Brown University, Rhode Island, USA, and is general partner of RKM & Associates. Since the start of his career in the mid 1960s, he has had experience in carbonate sedimentation and diagenesis and their application to petroleum exploration and reservoir characterization. Rob’s current interests center around the use of computer-based dynamic models in stratigraphic simulation.

Manuscript submitted September 30, 2003
Revised February 2, 2004
Accepted February 19, 2004
Press Version Proofread by Author June 4, 2004