Palaeogeography, Palaeoclimatology, Palaeoecology 441 (2016) 393–411

Contents lists available at ScienceDirect

Palaeogeography, Palaeoclimatology, Palaeoecology

journal homepage: www.elsevier.com/locate/palaeo

# Review: Short-term sea-level changes in a greenhouse world — A view from the Cretaceous

B. Sames [a,b,*], M. Wagreich [a], J.E. Wendler [c], B.U. Haq [d], C.P. Conrad [e], M.C. Melinte-Dobrinescu [f], X. Hu [g], I. Wendler [c], E. Wolfgring [a], I.Ö. Yilmaz [h,i], S.O. Zorina [j]

[a] University of Vienna, Department for Geodynamics and Sedimentology, Geozentrum, Althanstrasse 14, 1090 Vienna, Austria
[b] Sam Noble Oklahoma Museum of Natural History, 2401 Chautauqua Avenue, Norman, OK 73072-7029, USA
[c] Bremen University, Department of Geosciences, P.O. Box 330440, 28334 Bremen, Germany
[d] Smithsonian Institution, Washington DC, USA, and Sorbonne, Pierre & Marie Curie University Paris, France
[e] University of Hawaii at Mānoa, Department of Geology and Geophysics, School of Ocean and Earth Science and Technology, Honolulu, HI 96822, USA
[f] National Institute of Marine Geology and Geoecology (GeoEcoMar), Str. Dimitrie Onciul Nr. 23, 024053 Bucharest, Romania
[g] Nanjing University, School of Earth Sciences and Engineering, Hankou Road 22, Nanjing 210093, PR China
[h] Middle East Technical University, Department of Geological Engineering, 06531 Ankara, Turkey
[i] The University of Texas at Austin, Department of Geological Sciences, 2275 Speedway Stop C9000, Austin, TX 78712-1722, USA
[j] Kazan Federal University, Department of Paleontology and Stratigraphy, Kremlyovskaya str. 4/5, Kazan 420008, Russia

ARTICLE INFO

Available online 3 November 2015

*Keywords:*
Cretaceous greenhouse
Eustasy
Relative sea-level change
Aquifer-eustasy
Sequence stratigraphy
Orbital cycles

ABSTRACT

This review provides a synopsis of ongoing research and our understanding of the fundamentals of sea-level change today and in the geologic record, especially as illustrated by conditions and processes during the Cretaceous greenhouse climate episode. We give an overview of the state of the art of our understanding on eustatic (global) versus relative (regional) sea level, as well as long-term versus short-term fluctuations and their drivers. In the context of the focus of UNESCO-IUGS/IGCP project 609 on Cretaceous eustatic, short-term sea-level and climate changes, we evaluate the possible evidence for glacio-eustasy versus alternative or additional mechanisms for continental water storage and release for the Cretaceous greenhouse and hothouse phases during which the presence of larger continental ice shields is considered unlikely. Increasing evidence in the literature suggests a correlation between long-period orbital cycles and depositional cycles that reflect sea-level fluctuations, implying a globally synchronized forcing of (eustatic) sea level. Fourth-order depositional sequences seem to be related to a ~405 ka periodicity, which most likely represents long-period orbital eccentricity control on sea level and depositional cycles. Third-order cyclicity, expressed as time-synchronous sea level falls of ~20 to 110 m on ~0.5 to 3.0 Ma timescales in the Cretaceous, are increasingly recognized as connected to climate cycles triggered by long-term astronomical cycles that have periodicity ranging from ~1.0 to 2.4 Ma. Future perspectives of research on greenhouse sea-level changes comprise a high-precision time-scale for sequence stratigraphy and eustatic sea-level changes and high-resolution marine to non-marine stratigraphic correlation.


**Contents**

1. Introduction . . . 394
2. Fundamentals of relative and eustatic sea level and sea-level change . . . 395
   2.1. Sea level and sea-level fluctuations: classification and measurement . . . 395
   2.2. Timescales and amplitudes of sea-level change . . . 396
   2.3. Drivers and mechanisms of long- and short-term eustatic sea-level changes . . . 397
   2.4. Physico-chemical intrinsic contributions: ocean water temperature and salinity — steric sea-level change . . . 400
   2.5. The cryospheric contribution — glacio-eustasy . . . 401
   2.6. Continental water storage and release contributions . . . 401
   2.7. Solid-Earth contributions . . . 403
   2.8. Geoid contributions . . . 404
   2.9. Reconstructing sea-level changes in the geologic record . . . 404

* Corresponding author at: University of Vienna, Department for Geodynamics and Sedimentology, Geozentrum, Althanstrasse 14, 1090 Vienna, Austria.
*E-mail address:* benjamin.sames@univie.ac.at (B. Sames).

http://dx.doi.org/10.1016/j.palaeo.2015.10.045
0031-0182/© 2015 Published by Elsevier B.V.

2.10. Constructing short-term sea-level curves from the geologic record . . . 405
3. The Cretaceous world . . . 405
3.1. Cretaceous short-term sea-level changes and their drivers . . . 405
3.2. Cretaceous short-term eustatic changes as a stratigraphic tool. . . . 406
3.3. Cretaceous cyclostratigraphy and marine to non-marine correlations . . . 407
4. Conclusion and perspectives . . . 407
5. Author contributions . . . 407
Acknowledgements . . . 407
References. . . . 408

## 1. Introduction

Global warming and associated global sea-level rise resulting from steady waning of continental ice shields and ocean warming have become issues of growing interest for the scientific community and a concern for the public. Sea level constitutes a basic geographic boundary for humans and sea-level changes drive major shifts in the landscape. A global sea-level rise even on the scale of a meter or two could have major impact on mankind, particularly in vulnerable coastal areas and oceanic island regions (e.g. El Raey et al., 1999; Nicholls, 2010; Nicholls and Cazenave, 2010; Caffrey and Beavers, 2013; Church et al., 2013; Cazenave and Le Cozannet, 2014). Adaption strategies for vulnerable regions have thus become major concerns for maritime nations worldwide. Identified drivers of recent sea-level rise initiated by global warming are mainly (1) accelerated discharge of melt water from continental ice shields into the oceans; (2) thermal expansion of seawater (e.g. Cazenave and Llovel, 2010; Church et al., 2010); and (3) potential oceanic forcing of ice sheet retreat on ice shelves (e.g. as for parts of Antarctic and Greenland and ice sheets, see Alley et al., 2015).

However, the processes and feedback for sea-level change are highly complex. For example, the increasing temperature of the oceans and increased freshwater discharge into the oceans through melting ice shields can lead to disruptions and changes in the thermohaline ocean circulations (such as the shutdown or slowdown of the Gulf stream, e.g. Rahmstorf et al., 2015; Robson et al., 2014; Velinga and Wood, 2002) that are among the main drivers of global climate (e.g. Hay, 2013). At the same time, the magnitude of future sea-level rise remains highly uncertain (e.g. Nicholls and Cazenave, 2010; Church et al., 2013), and ocean circulation and climate models (coupled atmosphere–ocean general circulation models) are open to non-unique interpretations, making the topic controversial not only within the scientific community and its opinion leaders, but also among policy makers and the media. In addition, regional, non-climate related components of relative sea-level fluctuations (such as tectonically-induced and anthropogenic subsidence, isostatic compensation of increasing water load) further add to the complexity of the matter (e.g. Syvitski et al., 2009; Conrad, 2013).

To study sea-level changes over time, both today and in the sedimentary record, the main focus is on the globally synchronous changes, i.e. so-called eustatic sea-level changes — in contrast to relative or regional sea-level changes (termed *eurybatic* shifts by Haq, 2014, see Section 2.1 for details). The term *eustasy* goes back to the Austrian geologist Eduard Suess in 1888 who introduced the term “eustatic movements” for the globally synchronous sea-level changes preserved in the stratigraphic record, which is how it is used in the modern sense (for details see Wagreich et al., 2014; Şengör, 2015). In the context of eustatic sea-level change, terms such as “glacio-eustasy” or “glacio-eustatic sea-level changes” (eustatic sea-level changes caused by the waxing and waning of continental ice shields that lead to an increasing or decreasing water volume in the oceans), thermo-eustatic sea-level changes, tectono-eustatic sea-level changes etc., have subsequently been coined. However, all measures of sea-level change amplitude (rises and falls measured in meters) in any given region of the globe are always local (‘regional’ or ‘relative’ sea-level changes, see Conrad, 2013; Haq, 2014; Cloetingh and Haq, 2015), even when there is a strong underlying global signal since they are a product of both local vertical movements (solid-Earth factors) and eustasy (changes in ocean water volume and/or the volume of ocean basins, i.e. ocean capacity or “container volume”, respectively; refer to Section 2 for details). Consequently, eustatic sea-level amplitudes cannot be measured directly; quantitative estimates for amplitudes of past sea-level changes thus rely on averaged global estimates of eustatic changes in relation to a fix point, e.g. the Earth's center (see Haq, 2014).

Correlation, causes and consequences of significant short-term (cycles of 3rd and 4th order, i.e. about 0.5–3.0 Ma, and a few tens of thousands to ~0.5 Ma, respectively) sea-level changes which are recorded in Cretaceous sedimentary archives worldwide are addressed by the UNESCO-IUGS IGCP project 609 “Climate–environmental deteriorations during greenhouse phases: Causes and consequences of short-term Cretaceous sea-level changes” (http://www.univie.ac.at/igcp609/; lasting from 2013–2017). The project serves as a communication and collaboration platform bringing together specialists and research projects from around the world (from universities and other research facilities, from the industry and from stratigraphic consulting companies).

The Cretaceous (145–66 million years ago) was different from our present world in many respects, including climatic conditions (greenhouse world in general, with potential episodic glaciations, particularly during the Early Cretaceous), climate change patterns, oceanographic conditions and generally high global (eustatic) sea level. It was a time of enormous evolutionary changes, particularly on land, and critical to the origin and development of modern continental ecosystems. As the youngest prolonged greenhouse interval in Earth history, the Cretaceous constitutes a well-studied period in these respects (e.g. Hay, 2008; Hay and Floegel, 2012; Hu et al., 2012; Wagreich et al., 2014). The Cretaceous greenhouse period provides a suitable laboratory for better understanding of the causes and consequences of global short-term sea-level changes over a relatively long time interval with different (intermittently extreme) climates that may have important relevance for predictive models of future sea levels (e.g. Hay, 2011; Kidder and Worsley, 2012).

Our views of Cretaceous climates have changed during the last decades, from a warm, equable Cretaceous greenhouse to a Cretaceous that is subdivided into 3–4 longer-term climate states: a cooler Early Cretaceous greenhouse with the possibility of “cold snaps”, a very warm greenhouse mid-Cretaceous ("Supergreenhouse") including short-lived ‘hothouse’ periods with widespread anoxia and a possible reversal of the thermohaline circulation (HEATT episodes of ‘haline euxinic acidic thermal transgression’, see Kidder and Worsley, 2010; Hay and Floegel, 2012), and a Late Cretaceous warm to cool greenhouse evolution (e.g. Skelton, 2003; Kidder and Worsley, 2010, 2012; Föllmi, 2012; Hay and Floegel, 2012; Hu et al., 2012). Moreover, an increasing number of short-term climatic events within the longer-term trends are also reported (e.g. Jenkyns, 2003; Hu et al., 2012).

Cyclic sea-level changes and corresponding depositional sequences and sedimentary cycles are usually explained by the waxing and waning of continental (polar) ice sheets. However, though Cretaceous eustasy involves brief glacial episodes, for which there is evidence at least in the Early and the latest Cretaceous (e.g. Alley and Frakes, 2003; Price and Nunn, 2010; Föllmi, 2012), the presence of continental ice sheets during the remainder of the Cretaceous is controversial, and remains particularly enigmatic for the mid-Cretaceous extreme greenhouse period (Aptian to Turonian) with "hothouse" episodes and global average temperature maxima during the later Cenomanian to Turonian (e.g. Hay and Floegel, 2012).

For these reasons, IGCP 609 is focusing more on the causes and mechanisms of short-term eustatic sea-level changes in the mid-Cretaceous "Supergreenhouse" or "hothouse" periods (Cenomanian–Turonian) during which continental ice sheets are highly improbable and, thus, other mechanisms have to be taken into consideration to explain significant short-term eustatic changes, such as "aquifer-eustasy" (Jacobs and Sahagian, 1995; Hay and Leslie, 1990; Wendler and Wendler, 2016-in this volume; Wendler et al., 2016-in this volume) or "limno-eustasy" (Wagreich et al., 2014; see also Section 2.6.). The focus on short-term eustatic sea-level changes is also warranted because of their importance for stratigraphic applications: resulting marine depositional sequences and sequence boundaries would be synchronous and correlatable — the challenge, however, is proving their supraregional to global correlations at sufficient resolution. This crucial point is addressed by IGCP 609, i.e., the interrelation of short-term climate changes and eustatic sea-level changes, their analysis for astronomically driven cyclicities, and their cyclostratigraphic application.

Recent refinements of the geological timescale using new radiometric data and numerical calibration of bio-zonations, carbon and strontium isotope curves, paleomagnetic reversals, and astronomically calibrated timescales (for the latest Cretaceous) have made major advances for the Cretaceous. International efforts are improving the Cretaceous timescale to yield a resolution comparable to that of younger Earth history. It is now possible to correlate and date short-term Cretaceous sea-level records with a resolution appropriate for their detailed analysis (e.g. Wendler et al., 2014), that is to say, a resolution on Milankovitch astronomical scales (mainly in the band of 405 and 100 ka eccentricity cycles). With respect to the Cretaceous, orbital tuning and floating timescales have become available for the latest Campanian through Maastrichtian (see Ogg et al., 2012, and Batenburg et al., 2014) and are continuously being advanced backwards in stratigraphy. Respective correlations and precise ages of sequence boundaries and cycles not only provide an advanced tool for global correlations at high resolution, but also facilitate the testing of hypotheses concerning the interrelationships of astronomically forced climate events and cyclicities, corresponding sea-level fluctuations and their control and feedback mechanisms, such as the "aquifer- or limno-eustatic hypothesis".

Consequently, major objectives of IGCP 609 are: (1) to correlate high-resolution sea-level records from globally distributed sedimentary archives to the new, high-resolution absolute Cretaceous timescale, using marine carbonate isotope curves and orbital (405, 100 ka eccentricity) cycles. This will resolve the question of whether the observed short-term sea-level changes are regional (tectonic) or global (eustatic) and determine their possible relation to climate cycles; (2) to facilitate the calculation of rates of sea-level change during the Cretaceous greenhouse episode, and during its (mid-Cretaceous) Supergreenhouse period. Rates of geologically short-term sea-level change on a warm Earth will help to better evaluate recent global change and to assess the role of feedback mechanisms such as thermal expansion/contraction of seawater, subsidence of continental margins and adjacent ocean basins due to loading by water, changing vegetation of the Earth System, changes in the hydrologic cycle etc., as well as (3) to further investigate the relation of sea-level highs and lows to major climate-oceanographic events such as ocean hypoxia and oxidation events, as represented in the sedimentary archives by black shales and oceanic red beds, and the evaluation of the evidence for ephemeral glacial episodes or other climate events, i.e., whether or not specific sea-level peaks are associated with glacial episodes. Multi-record and multi-proxy studies are needed in order to develop a high-resolution scenario for sea-level cycles and allow the development of quantitative models for sea-level changes in greenhouse episodes.

In this introductory review, we give an up-to-date overview on the fundamentals and background of sea level and sea-level change with respect to research on "short-term climate and sea-level changes" and their interrelationship today and in the geologic (sedimentary) record, with focus on the Cretaceous greenhouse period. Herein we follow the "IUPAC-IUGS Recommendations 2011" (Holden et al., 2011) in the usage of units of time, i.e. that the same units (a = year, ka = 1000 years, Ma = 1 million years) are applied to express both absolute time and time duration.

## 2. Fundamentals of relative and eustatic sea level and sea-level change

### *2.1. Sea level and sea-level fluctuations: classification and measurement*

The terms "sea level", "relative sea level" and "relative sea-level change" have varied in their usage among different authors and across scientific research groups and disciplines over time, through the historical development of respective research (Shennan, 2015). As a result, there is not only ambiguity in the use of terms concerning how sea level can be "relative" – elevation relative to the Earth's surface or elevation relative to the present – but there are also differences between modern oceanographers and geologists regarding how different terms are used (e.g. Shennan et al., 2012; Shennan, 2015). While modelers have presented explicit definitions with mathematical notation and defined sea level "as the elevation of the geoid (mean height of the sea surface averaged over several decades) in relation to the solid surface of the earth" (Shennan, 2015, p. 6), this is called 'relative sea level' in common geological use (op. cit.). Another variation would be the consideration of "change" within relative sea-level change as process rather than a measurement difference (e.g. a 'sea-level shift') attributed to a specific cause (Shennan, 2015), such as the melting of continental ice shields. Here we follow the definitions from the "Handbook of Sea-Level Research" (reviewed by Shennan, 2015 therein) as given below.

In general, a distinction is drawn between two fundamental types of sea level or sea-level shifts (change), respectively: (a) relative, regional or "eurybatic" (after Haq, 2014) sea-level shifts on the one hand, and (b) global or eustatic sea-level shifts on the other hand. These two differ in the geographic dimension of their geologic record (and the possibility of detection), in their degree of synchronicity (particularly important in the analysis of the geological record), and in the way they can be measured or calculated. The following definitions apply (if not explicitly indicated, terms and processes given in this section will be elucidated in the subsequent section in detail):

A) *Relative (regional) sea level or sea-level change*, respectively: "For each geographical location and time, sea level is the difference between the geoid and the solid rock or sediment surface of the Earth, both measured with reference to the centre of the Earth" (Shennan, 2015, p. 7; cited without symbols for mathematic variables and corresponding equations; see also p. 8, Fig. 2.4. therein). Based on this definition, sea level equals the common geological usage of the term "relative sea level" (Shennan, 2015). Therefore, a sea-level change "is given by the change in sea surface height minus the change in solid surface height over the period of interest" (Shennan, 2015, p. 7). With these definitions it is apparent that there are different components to be considered when measuring sea level and calculating sea-level change: the water (volume) component and the solid-Earth component and their interrelationships (see Sections 2.3. and 2.6. for details). Consequently, Shennan and Horton (2002, p. 511), define relative sea level as the sum of global/eustatic sea level including ocean water and ocean basin ("container volume" or capacity) changes (the "time-dependent eustatic function"), glacial isostatic adjustment (total isostatic effect of the glacial rebound process of the lithosphere including the glacio-isostatic and hydro-isostatic load and unload contributions), tectonic effects (including active and passive thermal subsidence, effects of dynamic topography, e.g. Miller et al., 2011; Conrad, 2013), and local effects (such as sediment compaction and changes in tidal range).

Though not applicable to the pre-Quaternary time interval, it must be mentioned that there is another common convention to define a change in relative sea level for Quaternary and Holocene time scales: a definition as change relative to present sea level (Shennan, 2015).

B) In theory, *eustatic (global) sea level* "is the sea level that would result from distributing water evenly across a rigid, non-rotating planet and neglecting self gravitation in the surface load (Mitrovica and Milne, 2003, cited after Shennan, 2015, p. 6). Since Earth is not a rigid planet, and it does rotate and has self-gravitation, it is not possible to record eustatic sea level (and change) at any single locality on Earth (Shennan, 2015). Actually, all measurements of amplitudes of sea level or sea-level change (recent and past rises and falls measured or reconstructed in millimeters to meters) in any given region are always local, and consequently "relative" or "regional", even when there is a strong overlying global signal (Haq, 2014). In other words: Eustatic sea-level amplitudes and changes cannot be measured — these are averaged global estimates of eustatic changes in relation to a fix-point, for example the Earth's center (e.g. Haq, 2014). Corresponding to their respective drivers, different composite terms have been coined for eustatic sea-level changes, such as glacio-eustasy, aquifer/limno-eustasy, thermo-eustasy, and tectono-eustasy, the details of which are summarized in the Sections 2.3 and the following.

Regarding the reconstruction of sea level and sea-level changes from the geologic record, the differentiation of eurybatic (regional, or relative) and eustatic (global) sea-level changes (Fig. 1) and the respective proportion of each signal at a given locality or region is a critical issue (see e.g., Moucha et al., 2008; Müller et al., 2008; Conrad and Husson, 2009; Conrad, 2013; Haq, 2014), the disregard of which can lead to strong over- or underestimations of amplitudes (e.g., Miller et al., 2005a). We must also bear in mind that depending on the time interval in question, concerning the geologic record of deep time we can only detect and correlate *significant* (i.e. observable) sea-level changes of certain minimum amplitudes. The minimum of the latter, in turn, depends on the stratigraphic resolution available, which tends to decrease as we go back in time. These issues and the subject of how to reconstruct paleo-water depths and sea-level changes are overviewed in Section 2.8.

### *2.2. Timescales and amplitudes of sea-level change*

Sea level fluctuates at varying rates (timescales and amplitudes), geographically and over time. Analyzing and modeling currently available direct measurements (from tide gauges from different parts of the world: measurements available since about 1700 and without gaps since the 1860s; and satellite altimetry starting in 1993 with the TOPEX/Poseidon radar altimeter satellite, see e.g. Church et al., 2013; Mitchum et al., 2010; Woodworth and Menéndez, 2015), are usually made on annual, decadal and centennial timescales and exhibit amplitudes of few millimeters to a few meters. This also includes sea-level prediction and the impact of sea-level rise on mankind as well as feasible responses to it. Between 1900 and 2010 estimated global mean sea level rose by approximately 1.7 mm/year, accelerating to about 3.2 mm/year since the 1990s (e.g. Church et al., 2013; Hay et al., 2015; Mitchum et al., 2010; Woodworth et al., 2009; Woodworth and Menéndez, 2015; and references in these).

In contrast, detectable and calculable sea-level fluctuations in the geologic record exhibit different, normally longer, time intervals and

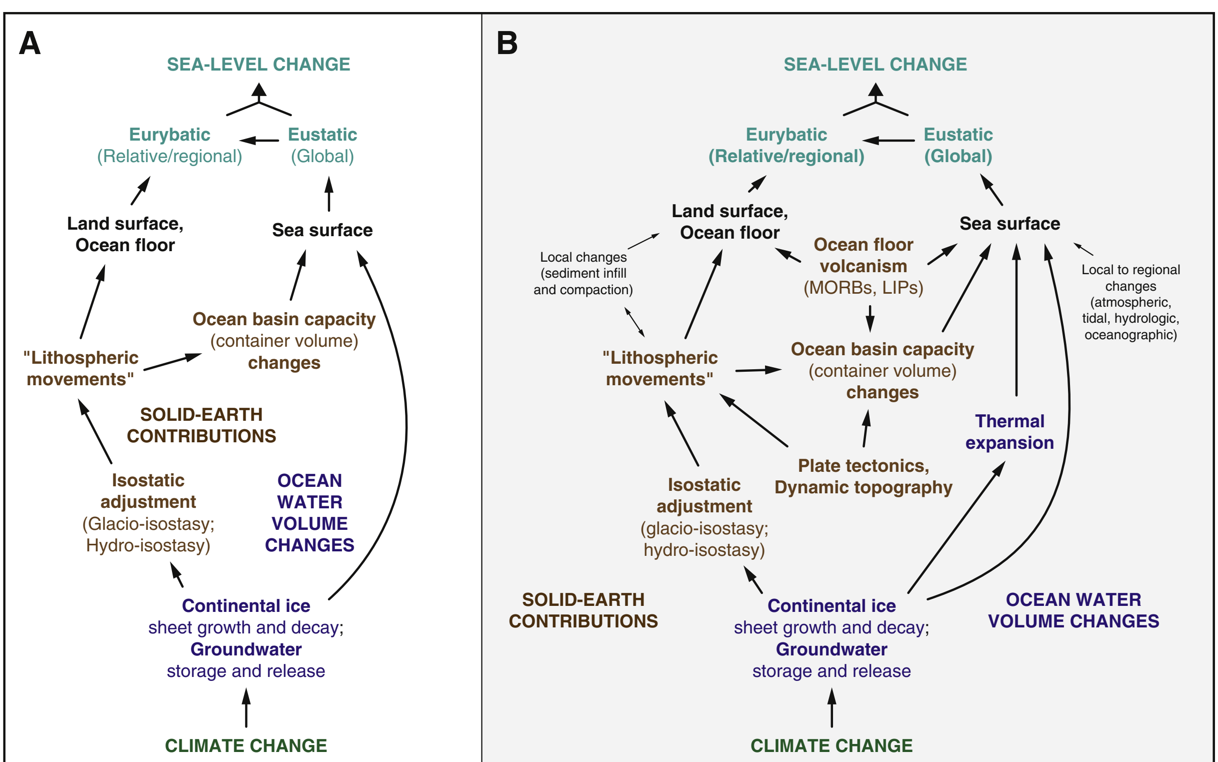


**Fig. 1.** Scheme of interrelationships of global-, regional and local-scale processes and factors that contribute to eurybatic and eustatic sea-level changes (strongly modified after Shennan, 2015), with focus on short-term effects (<3 Ma). (A) Simple relationship scheme. "Lithospheric movements" comprising all tectonic-related plate movements. B) Complex relationship scheme. Note that geoid contributions are not included. Abbreviations: MORBs — Middle ocean ridge basalts; LIPs — Large igneous provinces (here submarine basalt plateaus).

larger amplitudes due to preservational bias in the depositional record. The main limiting factors are strongly dependent on the severity of each sea-level event (i.e. were these sea-level changes, with their given amplitudes and geographic scales, actually observable in the geologic record), the stratigraphic resolution available (i.e. did these fluctuations occur within time frames observable and correlatable in the geologic record), the definition and interpretation of sequence boundaries, and the driving factors and mechanisms of respective sea-level changes.

In the Cretaceous we are mainly dealing with significant global (eustatic, see Section 2.3 for details), short-term cyclic sea-level fluctuations of about 0.5–3.0 Ma (so-called 3rd-order cyclicities) and a few tens of thousands to about 0.5 Ma (so-called 4th-order cyclicities) duration (e.g. Haq, 2014). The 405 ka cyclicity (coeval with the long spectral components of orbital eccentricity) appears to be a prevalent signal and fundamental feature of sedimentary sequences throughout the Phanerozoic (Gale, 1996; Gale et al., 2002; Gradstein et al., 2012; Haq, 2014, and references therein). Estimated amplitudes (averaged global estimates, see Section 2.1. above) of Cretaceous eustatic (global and time synchronous) sea-level changes (3rd order) greatly vary and are in the order of about 20–110 m (Haq, 2014). The recorded long-term trends (2nd-order cyclicity, >5 to ~100 Ma) exhibit changes within a few tens of meters range during the Cretaceous, and global sea-level is considered to have been between ~65 and 250 m higher than the present day mean sea level (Haq, 2014).

### 2.3. Drivers and mechanisms of long- and short-term eustatic sea-level changes

Sea-level changes result from a complex combination and interrelationship of operative mechanisms, processes, and influencing factors that are different in modality, magnitude, extent, and timescale. These can modify regional and/or global sea level, and differ in their eustatic contribution to the local/regional sea-level signals (see Figs. 1 to 5).

In principle, fluctuations in eustatic sea level are caused by two major categories of mechanisms, which can be grouped into acting on either "long-term" or "short-term" scales (see Section 2.2). Fluctuations in global eustatic sea level originate from (A) changes in the total available volume of ocean/marine basins ("container volume"), and (B) changes in the cumulative volume of water in the oceans (ocean–continent and ocean–mantle water distribution).

A) *Processes related to changes in the volume of ocean/marine basins:* The first group of mechanisms leads to changes in the volume of ocean basins (capacity or "container volume") and comprise shape and size changes (various processes) of ocean basins, their sedimentary or magmatic filling (recurrent periods of submarine volcanic pulses: ocean ridge basalts, syn-rift volcanism), and "dynamic topography" (see below). These processes cause net contractions or expansions of the ocean basins, which in turn causes sea-level rises or sea-level falls, respectively. Related processes and effects on sea-level change are mainly interconnected solid-Earth driven ones, and mostly act on longer scales, i.e., 2nd to 1st-order 'cycles' in the ranges of several (>5) Ma to over 100 Ma (e.g. Conrad, 2013; Cloetingh and Haq, 2015). Sea-level changes based on processes related to tectonic movements of the Earth's plates are referred to as tectono-eustatic sea-level changes (and the process as tectono-eustasy). Related processes are: (1) ocean floor volcanic activity, i.e. (1a) ocean crust production at mid-ocean ridges (changes can displace sea water equating to a few hundreds of meters eustatic sea-level changes within ~100 Ma, Pitman, 1978; Kominz, 1984; Xu et al., 2006; Müller et al., 2008; Conrad, 2013) and (1b) eruption of large igneous provinces (which can displace enough water to create ~100 m of eustatic sea level change, Harrison, 1990; Müller et al., 2008); (2) net changes in the areal extent of the oceans caused by continental orogeny or extension (which can create ~10s of meters of eustatic change, Kirschner et al., 2010); and (3) net subsidence or uplift of the ocean basins by mantle dynamics (changes to this "dynamic topography" can cause eustatic changes up to 1 m/Ma sustained over several 10s of Ma, Gurnis, 1990, 1993; Conrad and Husson, 2009; Spasojevic and Gurnis, 2012). Adding to these is sediment infill (sediment supply) from erosion of continental surfaces not covered by oceans, which also displaces enough sea water to cause up to ~100 m of eustatic sea-

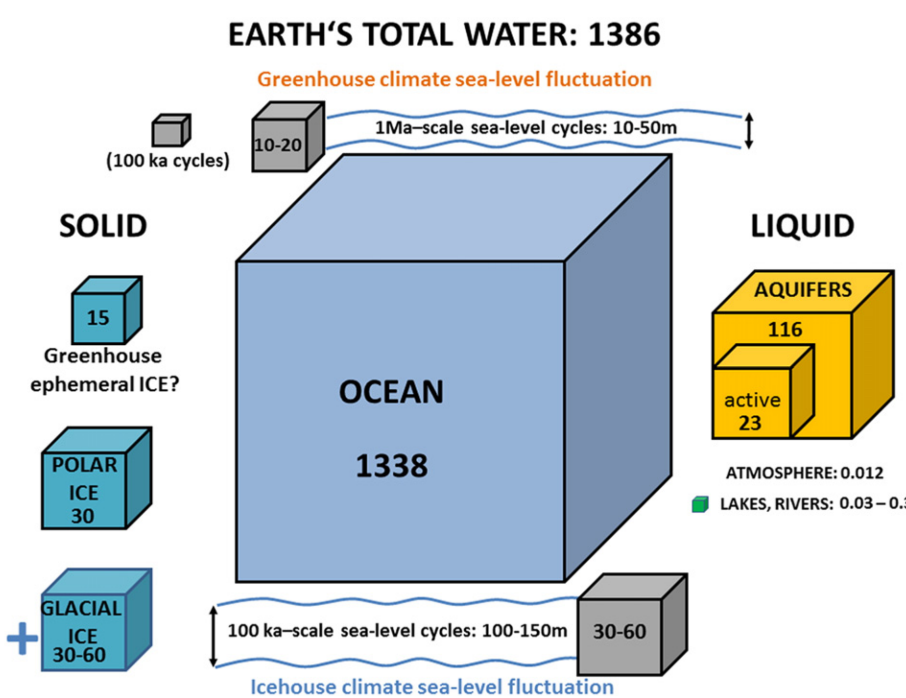


**Fig. 2.** Water volumes in the Earth system based on estimates by Hay and Leslie (1990). Water volumes are given in million $km^3$ for the ocean (blue box), solid ice (cyan boxes), continental aquifers (yellow boxes), surface water (green box) and the atmosphere (corresponding box is too small to plot). General average eustatic sea-level change values and respective water volumes (grey boxes) are given for greenhouse (top) and icehouse (bottom) climate states.

| MECHANISMS | Operative timescales | Water volume equivalents | Orders of magnitude in eustatic sea level | Potential extent | Cretaceous relevance |
|---|---|---|---|---|---|
| *Changes in ocean water volume* | | | | | |
| Thermal expansion (thermo-steric effect) | 1 to 10.000 a | ~ 0.2 × $10^6$ km³/1°C | ~ 5 to 10 m | Global | ? (*) |
| Continental glaciations/deglaciations | <0.01 to 0.1 Ma | ~ 25 × $10^6$ km³ | ~ 50 m to 250 m | Global | ? |
| Continental water storage and release | <0.01 Ma | 24 to 30 × $10^6$ km³ | 10 to 50 (80?) m | Global | ? |
| Water exchange with (deep) mantle | 0.1? to 1 Ma/1 Ga | Unknown | Unknown | Global? | moderate |
| *Changes in container volume (capacity) of ocean basins* | | | | | |
| GIA: Elastic rebound of lithosphere | instantaneous | n/a | Up to 100 m | Regional | ? |
| GIA: Viscous mantle flow | 0.0001 to 0.1 Ma | n/a | Up to 100 m | Regional | ? |
| Mean age of oceanic crust | 50 to 100 Ma | n/a | 100 to 300 m | Global | high |
| MORB production rate changes | 50 to 100 Ma | Up to 30 × $10^6$ km³ | 100 to 300 m | Global | high |
| Ocean floor volcanism (LIPs) | 1(0) to 10(0) Ma | n/a | 50 to 100 (500? to 1000?) m | Global | high |
| Mantle/lithosphere interactions | 1(0) to 10(0) Ma | ? | 10 to 30/100 (<1000) m | Regional/global | high |
| Intraplate deformation | 1(0) to 10(0) Ma | ? | 10 to 30/100 (<1000) m | Regional/global | high |
| Dynamic topography | >5 (10 to 100) Ma | ? | 100 to 300 (<1000) m | Regional/global | high |
| Sediment infill | 50 to 100 Ma | ~ 25 × $10^6$ km³ | 50 to 100 m | Global | low |

**Fig. 3.** Overview of mechanisms influencing regional (local/relative or eurybatic) and global (eustatic) sea levels and sea-level changes and their operative timescales, equivalent water or water displacement volumes, respectively, and the orders of magnitude of corresponding sea-level changes, potential extent of related sea-level changes, and considered relevance of each respective mechanism to the Cretaceous period (modified from Cloetingh and Haq, 2015; compiled including data from Jacobs and Sahagian, 1993; Miller et al., 2011; Hay and Leslie, 1990; Dewey and Pitman, 1997; Conrad, 2013). All these estimates are continuously debated and remain object to change to different degrees. Values and value ranges given are estimations for recent and geologic times, except for water volume equivalents for "continental glaciations/deglaciations" and "continental water-storage and -release", which are recent estimates only (cf. Chapter 2 also). The amplitude estimation for "continental glaciations/deglaciations" considers the whole Phanerozoic. Operative timescales column: a) All climate-related changes in ocean water volume (thermal expansion, continental glaciation/deglaciation, continental water storage and release) could operate on much longer timescales as well (10–100 Ma) as climate also fluctuates on those timescales (transition/changes between climate modes); b) The solid-Earth components given (mantle/lithosphere interactions, intraplate deformation, dynamic topography) are for eurybatic sea level, higher values in the operative timescales-column (add additional cipher in parentheses) for eustatic sea level; c) Elastic rebound of lithosphere is instantaneous and relevant timescales are the rate of mass loading, which are associated with climate change (decades or Milankovitch-cycle scales). Abbreviations for units of time follow the "IUPAC-IUGS Recommendations 2011" (Holden et al., 2011) in that the same units (a = year, ka = 1000 years, Ma = 1 million years) are applied to express both absolute time and time duration. Magnitude of eustatic sea-level is in meters (m). Abbreviations: GIA — Glacial Isostatic Adjustment; MORB — Middle Ocean Ridge Basalts; LIP — Large Igneous Provinces (here submarine basalt plateaus). (*) insufficient temporal resolution.

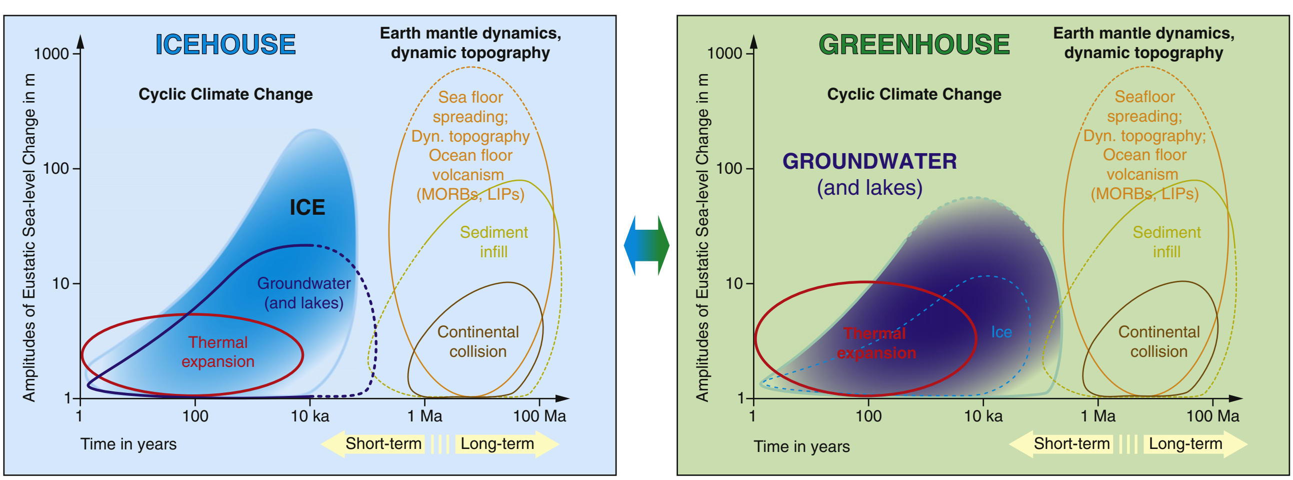


**Fig. 4.** Comparative log-scale diagram sketches of the timing and amplitudes of major geologic mechanisms for driving eustatic sea-level changes during icehouse (left) and greenhouse (right) climate modes, respectively (modified from Miller et al., 2005a; based on data from various authors including, among others Hay and Leslie, 1990; Jacobs and Sahagian, 1993; J. Wendler, unpublished; Wendler and Wendler, 2016-in this volume; see also Figs. 1, 2 and 3 herein). The focus here is on short-term processes in relation to cyclic climate change (3rd- to 4th-order cycles). Note that these diagrams are rough sketches to illustrate (1) eustatic sea-level change efficacy (amplitude) of selected factors relative to each other in the two different climate regimes, and (2) ranges of their main relevance in the geologic record (timing vs. amplitude), i.e. at short-term (4th- and 3rd-order cycles) or long-term (2nd-order cycles) scales. These are sketches intended to give important dimensions of mechanisms and processes, not to be read as a true graphical representation of measured or calculated data (in which case all components also would have to start at the point of origin). Dashed lines give dimensions of efficacy that are of lesser relevance in the Cretaceous. The dominant processes for short-term eustatic sea-level change are glacio-eustasy during icehouse phases on the one hand and aquifer-eustasy during warm greenhouse phases on the other hand.

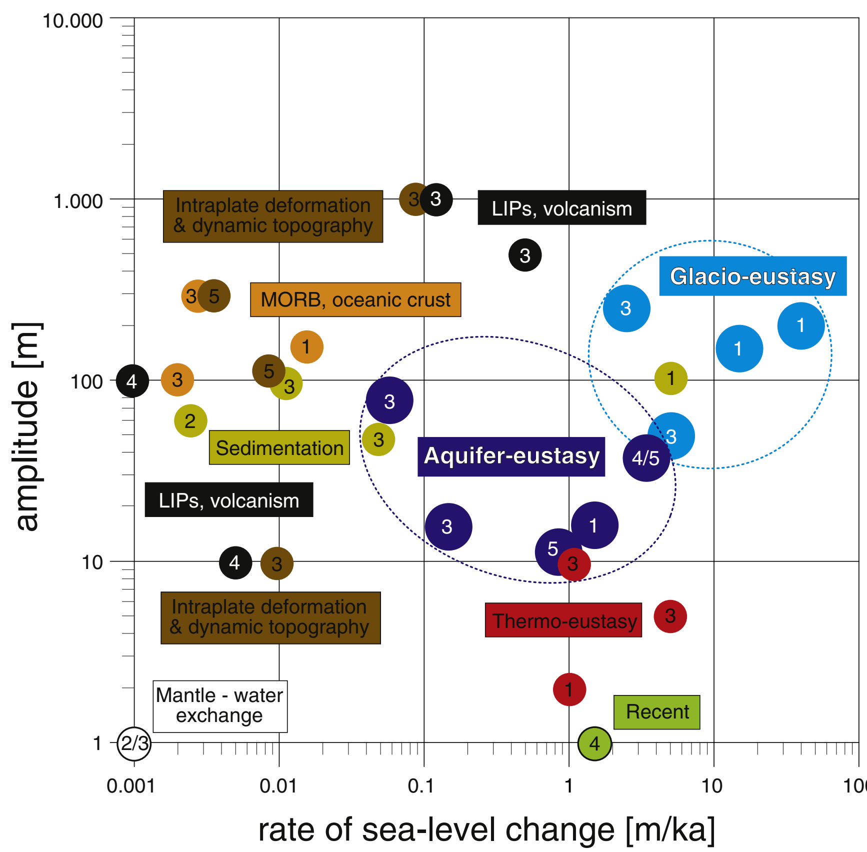


**Fig. 5.** Log-scale diagram of timing vs. rates (with minima and maxima if available from the same authors) of sea-level changes as inspired by a figure of Matt Hall (2011 in the "Agile Geoscience" Blog, http://www.agilegeoscience.com/blog/2011/4/11/scales-of-sea-level-change.html, accessed 08-10-2015). Rates calculated from respective sea-level amplitudes divided by duration and converted to meters per 1000 years, based on data compiled after Emery and Aubry (1991: 1), Conrad (2013: 2) and Cloetingh and Haq (2015: 3), herein: 4, Wendler and Wendler (2016-in this volume: 5). Overlapping circles have the same values (center of overlap). "Recent" refers to an average of the last 20–30 years.

level change (Harrison et al., 1981; Müller et al., 2008; Conrad, 2013).

In recent years, a complex of processes and feedbacks under the labels "dynamic topography" and "inherited landscapes" have received much attention as they affect local measurements of sea level and past reconstructions (see Cloetingh and Haq, 2015). We have learned that some of the processes mentioned above can refashion landscapes only regionally, and that solid-Earth processes are responsible for retaining lithospheric memory and its surface expressions (Cloetingh and Haq, 2015). Dynamic topography is vertical deflection of Earth's surface supported by stresses associated with mantle flow (e.g., Hager et al., 1985), with elevated topography above mantle upwelling and depressed topography above mantle downwellings (e.g. Flament et al., 2013 and references therein). Dynamic topography can change with time, as either mantle dynamics evolve or continents move laterally over different parts of the mantle (Gurnis, 1990, 1993; Spasojevic and Gurnis, 2012), inducing uplift or subsidence of the solid-Earth surface that affects both land- and seascapes. Under the term "inherited (regional) topography/landscapes" we subsume the effects of solid-Earth driven processes that lead to dynamic change in surface topography, for which dynamic topography is considered an important factor (see Section 2.7.). This process leads to net dynamic uplift of the seafloor by mantle flow (Conrad and Husson, 2009), and may also induce lateral variations in sea-level change by locally deflecting the ground surface (Conrad, 2013; Moucha et al., 2008). Surface topography reacts dynamically to both isostasy and mantle flow, resulting from lithospheric memory retained at various temporal and spatial scales (Cloetingh and Haq, 2015). Combinations of these processes can amplify, accelerate, cancel out, or decelerate each other. As we have learned recently it is essential within the scope of understanding sea-level changes to take these processes into consideration since they affect local measures of sea level, and thus, estimates of eustatic sea levels and sea-level changes as well, even on short-term timescales (sediment infill, dynamic topography, e.g. Conrad, 2013; Haq, 2014; Cloetingh and Haq, 2015). As Cloetingh and Haq (2015, p. 1258375-10) aptly put it "the interdisciplinary dissension between solid-Earth geophysics and soft-rock geology was at least partly due to the prevalent view within the sedimentologic community that post-rift tectonic processes are normally too slow to contribute to punctuated stratigraphy."

Nevertheless, basin volume changes resulting from solid-Earth processes (rock deformation, tectonics, volcanism, sedimentation, and mantle convection) occur on all timescales (Conrad, 2013). The critical point is whether or not the resulting effect on eustatic sea-level change is significant in the sense of: (1) being recognizable in the geologic record (and not wiped-out by erosional or other processes) and (2) significant in comparison to corresponding processes operating on a respective timescale. The Cretaceous, for example, represents a major episode of oceanic crust production that led to long-term sea-level rise and the eustatic sea-level highstands estimated between 170 and 250 m above today's sea level (e.g. Müller et al., 2008; Conrad, 2013; Haq, 2014), see Section 3.

On longer timescales (100 s of Ma, e.g. across supercontinental cycles and longer), the imbalance of water exchange with(in) the deep mantle (or "water sequestration within the mantle") may contribute significantly to eustatic sea-level fluctuations

(Kasting and Holm, 1992; Crowley et al., 2011; Korenaga, 2011; Sandu et al., 2011; Conrad, 2013). Sea-level rise (or fall) by this process results from imbalance in the rate of water exchange with the deep mantle by increased (or decreased) outgassing of the mantle (water release into the surface environment from melting of hydrated minerals in mantle rocks by degassing at mid-ocean ridges) or slower (or faster) loss of water into Earth's interior via subduction (i.e. water storage in hydrated minerals of the seafloor and their subduction into the deep mantle). However, Cloetingh and Haq (2015) discuss the possibility of water exchange with the mantle for explaining Cretaceous 3rd-order cycles, provided that the necessary leads and lags of water movements *within* the mantle can be demonstrated. In summary, the process of water exchange with the mantle is as yet not well understood with respect to operative timescales and dimensions of their sea-level affecting imbalances.

B) *Processes related to changes in the ocean's water volume*:
The second group of processes, predominantly governing short-term sea-level changes over much of Earth's history (but see Conrad, 2013, p. 1033 for the Cenozoic), concerns changes in ocean water volume. These processes include (1) the thermal expansion of sea water (thermo-eustatic sea-level changes, thermo-eustasy); (2) water storage and release (also "sequestration") on land as ice (i.e. the waxing and waning of continental ice sheets; glacio-eustatic sea-level changes, glacio-eustasy), and the imbalance in groundwater and lake water storage and release (aquifer-eustatic sea-level changes, aquifer-eustasy); and (3), potentially, imbalances (short-term leads and lags) in water exchange with the Earth's mantle as favored by Cloetingh and Haq (2015) (see previous section). These short-term processes act on 3rd- (about 405 ka to 0.5–3.0 Ma) to 4th-order scales (few tens of thousands to about 500 ka), and are mainly climate driven and cyclic. The interrelationship of astronomically forced climate cycles, which control short-term sea-level changes as well as (cyclic) variations in sediment deposition, is fundamental to geosciences, particularly to sequence- and cyclostratigraphy, and of central interest within the scope of IGCP 609. Therefore, these processes are elucidated in the following chapters.

### 2.4. Physico-chemical intrinsic contributions: ocean water temperature and salinity — steric sea-level change

The Earth's oceans exert a major control over the climate system since they store and transport huge quantities of heat (e.g. Broeker, 1991; Church et al., 2010; Hay, 2013; Rose and Ferreira, 2013). Understanding variation in the ocean's heat content in space and time is thus critical to our comprehension of the ocean's structure and circulation as well as its impact on climate variability and change (e.g. Church et al., 2010; Piecuch and Ponte, 2014). In addition, temperature changes in ocean water lead to heat induced thermal (volume) expansion or contraction. The amount of expansion depends on the quantity of heat absorbed, the initial water temperature (greater expansion in warm water), pressure (greater expansion at higher depth), as well as, to a smaller extent, salinity (greater expansion in water with higher salinity) (Church et al., 2010). Thus, temperature changes in ocean water contribute to global and regional sea-level change as an intrinsic factor: "a 1000-m column of sea water expands by about 1 or 2 cm for every 0.1 °C of warming" (Church et al., 2010, p. 143). 1 °C warming of the Earth's oceans is estimated to cause a eustatic sea-level rise of about 0.70 m (Conrad, 2013; Miller et al., 2009). Based on the estimated total volume of today's Earth ocean water of about $1335 \times 10^6$ km$^3$ (e.g. Hay and Leslie, 1990 and references therein), this would about equal a water volume of roughly $0.2 \times 10^6$ km$^3$ per 1 °C temperature change (depending on the initial temperature, depth and salinity, see above). Both temperature and salinity contributions, or their combined impact on density and volume, are significant for regional (relative) sea-level changes, while the temperature contribution is the dominant factor controlling global sea-level changes (Church et al., 2010).

The temperature and salinity effect on sea-water density and volume is called "steric effect" controlling the "steric sea level" or "steric sea-level changes", and correspondingly, the terms "thermosteric" (temperature contribution) and "halosteric" (salinity contribution) are used (e.g. Church et al., 2013). Along with glacier melting, ocean thermal expansion, i.e. global thermosteric sea-level rise, has been a major contributor to 20th century sea-level rise (together explaining 75% with high confidence excluding Antarctic glaciers peripheral to the ice sheet; the continental ice sheet contribution, i.e. Greenland and Antarctica, was smaller in the 20th century but has increased since the 1990s), and is projected to continue during the next centuries (Church et al., 2010, 2013; Piecuch and Ponte, 2014). Uncertainties in simulated and projected steric regional and global sea level remain poorly understood, and accordingly projected thermosteric sea-level rises based on climate models vary considerably (Church et al., 2013; Hallberg et al., 2013).

The physical steric effects, particularly the dominant thermosteric effect on sea-level change, were operating in the same way during Earth history. Indeed, in the geologic literature the term thermosteric sea-level (change) is substituted by thermal expansion or thermo-eustatic sea level (change). However, the thermosteric or thermo-eustatic effect and its contribution to sea-level change is even more difficult to calculate and model in deep time, as this requires detailed information not only on sea-water volumes, temperatures and salinity, but also on the variation of heat content and heat exchange in the oceans, changes in ocean mass from changes in ocean salinity, and past ocean circulations. In the Cretaceous, for example, the climate, continental distribution patterns and ocean circulations (thermohaline circulation) were significantly different (e.g. Friedrich et al., 2008; Hay, 1996, 2008; Hay et al., 1997; Hay and Floegel, 2012; Hasegawa et al., 2012). Moreover, as we can only estimate global (eustatic) sea-level changes from "measures" (which are estimates as well, cf. Section 2.9) of relative sea-level changes in the geologic record and discuss potential major controlling factors, we cannot make reliable estimates on the proportion of each respective factor of contribution to the total eustatic sea-level change.

In the geologic record, the differentiation of the thermosteric contribution from the cryospheric (see Section 2.5.) or continental water storage and release contribution (see Section 2.6.) is difficult because one of the main tools to estimate paleotemperatures and salinities of seawater, stable oxygen isotope fractionation and resulting isotope ratios ($\delta^{18}$O), likewise depends on temperature and salinity changes, and $\delta^{18}$O of sea water is directly affected by inflow of isotopic lighter ground- and melt-water. This issue becomes even more complex when differences in the oxygen isotope fractionation process and its net effect on sea water $\delta^{18}$O values during greenhouse climate modes are considered (see Wendler et al., 2016-in this volume; and Sections 2.5 and 2.6).

In addition, operative timescales and corresponding eustatic sea-level amplitudes resulting from volume changes of water in the oceans by thermal expansion or thermo-eustasy are in the range of 0.8–1.4 mm per year today (observed; modeled 0.97–2.02 mm per year; e.g. Church et al., 2013 given for the period 1993–2010; see also Church et al., 2010 and references therein), up to 10 m per thousand years (Miller et al., 2011) with total amplitudes estimated at between ~5–10 m (Jacobs and Sahagian, 1993; see also Fig. 3). Consequently, the contribution of thermo-eustatic sea-level changes to the total eustatic sea-level variation, though adding to it, is of lesser importance in the geologic record since it cannot be resolved. The thermo-eustatic sea-level changes have operative timescales that are several orders of magnitude smaller than the maximum stratigraphic resolution available for the Cretaceous (~20 ka), and their amplitudes

(~5–10 m, i.e. ≪25 m) are well within error ranges measured and estimated for short-term sea-level changes from the Cretaceous geologic record (mostly 25–75 m, e.g. Haq, 2014).

### 2.5. The cryospheric contribution — glacio-eustasy

Significant quantities of freshwater that can contribute to eustatic sea-level changes by changing the ocean water volume or its chemistry through inflow of meltwater (or freshwater removal by storage as continental ice, respectively) are stored in the continental ice sheets, most notably on Antarctica and Greenland, today (e.g. Steffen et al., 2010). Altogether, the present day cryosphere, i.e. ice sheets, ice caps, glaciers, and subsurface continental cryosphere (permafrost) on the continents contain an estimated water volume of about $24–30 \times 10^6$ km$^3$ (e.g. Hay and Leslie, 1990; Gleick, 1996) that is equivalent to ~64 m of sea-level and, applying isostatic compensation (of the water load by the crust and mantle), correlates to 45–50 m of eustatic sea-level rise for an ice-free world (Conrad, 2013). This estimate, of course, excludes oceanic floating ice (such as at the northern polar regions and floating glaciers peripheral to the continental ice sheets) because these have already displaced ocean water equal to the volume of water that would be created by their melting (hydrostatic equilibrium).

The waxing and waning of continental ice shields was certainly the dominant process relevant for eustatic short-term sea-level changes during the Holocene, and has been for much of the Earth history (e.g. Miller et al., 2011) during icehouse climate periods (however, for the past 40–50 years it has been outpaced by thermal expansion, e.g. Church et al., 2013). Resulting high-amplitude, rapid sea-level changes are called glacio-eustatic and operate at rates of up to 40 mm and more a year (during melt water pulses, Gehrels and Shennan, 2015; Miller et al., 2011), on timescales between 10 and 100 thousand years, and at amplitudes of 50 to 250 m (e.g. Conrad, 2013; Cloetingh and Haq, 2015; see Figs. 3, 4). During Snowball Earth times of the Precambrian (between ~780 and 630 Ma), i.e. for the hypothetic case that most or all continents were covered by ice sheets, a maximum of more than 600 m of sea-level fall has been modeled (Liu and Peltier, 2013).

However, for periods in Earth history where large continental ice sheets are considered to have been absent or highly improbable (warm greenhouse and hothouse intervals, e.g. much of the Cretaceous), the probability of continental ice as the only reservoir for significantly changing the ocean water volume was challenged in the early 1990s by the notion that climate controlled periodic continental groundwater storage and release may be an alternative mechanism for short-term sea-level changes instead of ice (Hay and Leslie, 1990; Jacobs and Sahagian, 1993). This idea has been revived especially for the Cretaceous by Wendler et al. (2011) and Föllmi (2012), and is currently tested and substantiated by these authors and other researchers (Wendler et al., 2014; Wendler and Wendler, 2016-in this volume; Wendler et al., 2016-in this volume; Wagreich et al., 2014), as discussed in Section 2.6.

A proxy to identify and calculate ice-volume and freshwater inflow changes in past oceans involves stable oxygen isotope rate changes over time, expressed as changes in sea water $\delta^{18}$O values. Based on isotope fractionation between the stable isotopes $^{16}$O and $^{18}$O during successive evaporation (preferring the lighter isotope) and condensation (preferring the heavier isotope) cycles, continental ice sequesters $^{16}$O and sea water becomes enriched in $^{18}$O during cold climates. Consequently, ice volume (and corresponding eustatic sea-level) changes can be reconstructed using marine carbonate $\delta^{18}$O values, mainly calcite tests of deep-sea benthic foraminifera (e.g. Shackleton and Kennett, 1975). Oxygen isotopes in marine sediments vary with periods that mirror orbital Milankovitch cyclicity, and constitute an important proxy for deciphering Quaternary cycles (e.g., Hays et al., 1974). During the Pleistocene, ice volume controlled two-thirds of the measured variability in oxygen isotope records, while temperature variations accounted for the other one-third (Miller et al., 2011). Thus, cyclic changes in stable oxygen isotope ratios connected to sea-level changes were used also to argue for glacio-eustasy in deep-time (e.g., Miller et al., 2005a, 2005b).

However, the use of oxygen isotopic ratios as an ice volume proxy is not straightforward and has many complications discussed in detail by Haq (2014). His conclusion was that although bulk carbonate isotopic curves could be used for estimating relative magnitudes of eustatic variations and aid us in determining the timing of eustatic events, they cannot be used as a quantitative measure of ice volume changes in deep time (Haq, 2014). Beyond this, the respective climate modes need to be more strongly considered for the interpretation of eustatic sea-level changes from shifts in seawater $\delta^{18}$O values. Thus far, usual reasoning equates positive shifts in seawater $\delta^{18}$O values with cooling and increasing continental ice volumes, which, in turn, correspond to eustatic sea-level falls that would be correlated with regressions (regressional cycles) in the geologic record or sequence stratigraphic interpretations. However, based on evidence from Cretaceous data, Wendler et al. (2016-in this volume) and Wendler and Wendler (2016-in this volume) present a new, more sophisticated interpretation of the differences in the oxygen-isotope fractionation process between icehouse and greenhouse (plus "hothouse") climate modes. Based on the assumption that glacio-eustasy dominates oxygen-isotope fractionation during icehouse conditions whereas aquifer-eustasy (see Section 2.6) is dominant during greenhouse conditions, Wendler and Wendler (op. cit.) discuss the corresponding differences in the effects of temperature and continental water volume on oxygen-isotope fractionation and the resulting net effects on seawater $\delta^{18}$O values. Following these authors (Wendler and Wendler, 2016-in this volume) the climate mode has considerable impact on paleoceanographic and paleoclimatic interpretations based on seawater $\delta^{18}$O values. Wendler and Wendler (op. cit.) present arguments and data that can explain positive shifts in seawater $\delta^{18}$O values and their correlation to high sea levels and transgressions, not regressions as previously thought, during the middle and late Turonian greenhouse climate.

Another important regional side effect of growth and decay of continental ice sheets (or continental groundwater reserves, see Section 2.6.) on short timescales is glacial isostatic adjustment (GIA), i.e. the isostatic rebound of the lithosphere during (ongoing melting process or groundwater release) and subsequent to continental ice (or continental groundwater, see Section 2.6.) load removal, particularly along the continental margins and adjacent ocean basins (e.g. Farrell and Clark, 1976; Mitrovica and Peltier, 1991; Milne and Mitrovica, 1998, 2008; Mitrovica and Milne, 2003). This is a solid-Earth contribution that operates on timescales of tens of thousands of years, and includes both the melting ice (glacio-isostatic) and (ground-)water (hydro-isostatic) load contributions (Shennan and Horton, 2002, p. 511), which affects relative/local sea-level measures (refer to GIA: glacial isostatic adjustment in Section 2.7 for details, and Fig. 3).

### 2.6. Continental water storage and release contributions

Continents provide the main storage capacity to effectively remove water from the oceans, with considerable potential to affect global sea level by changing ocean water volume (e.g. Hay and Leslie, 1990; the amount of water that can be stored in the atmosphere is negligible for affecting global sea level change, see Figs. 2, 3, and 4 for orders of magnitude/proportions). Apart from major ice shields, the only other significant water reservoirs on the continents are lakes and (much more important as to storage capacity) aquifers, i.e. porous sediments that may fill up with groundwater (see Fig. 2). Particularly during periods in Earth history where large continental ice sheets are considered to have been absent or highly improbable (warm greenhouse and hothouse intervals, e.g. much of the Cretaceous), the hypothesis that ice would be the only possible way of significantly changing the ocean water volume was challenged in the early 1990s by considerations that climate-controlled periodic continental groundwater storage and release could have contributed the major component to short-term sea-level changes instead of ice (Hay and Leslie, 1990; Jacobs and Sahagian, 1993). The groundbreaking idea was that for todays' Earth the calculated 'available' or 'active' groundwater volume (for being added to, or released from, the continents, thus affecting sea-level) would approximately equate to the water volume stored in continental ice shields, whereas the overall water capacity of lakes and rivers is almost negligible proportionally (Fig. 2; Hay and Leslie, 1990). Since then, this idea has been revived with particular focus on the Cretaceous, namely by Wendler et al. (2011) and Föllmi (2012), and is currently being tested and substantiated by these authors and other researchers (Wendler et al., 2014; Wendler and Wendler, 2016-in this volume; Wendler et al., 2016-in this volume; Wagreich et al., 2014), as discussed below. Consequently, this led to the hypothesis of “groundwater-driven eustasy”, termed “aquifer-eustasy” (see Hay and Leslie, 1990; Jacobs and Sahagian, 1993, 1995; Wendler et al., 2011, 2014; Wendler et al., 2016-in this volume) or “limno-eustasy”, alternatively (Wagreich et al., 2014; but see the subsequent paragraphs for details).

The fundamentals of the hypothesis of groundwater-driven eustasy go back to Hay and Leslie (1990, and references therein) who, based on estimates of pore space in continental sediments and their water-bearing potential, calculated the total available pore space and water capacity of surface and subsurface aquifers within continental blocks ($50.8 \times 10^6$ km$^3$), the subsurface aquifers of which being the major reservoir because they provide by far the major storage capacity. These authors also differentiated between sediments lying below sea level, which constitute the major part, and storage capacity that is permanently saturated with water (and, thus, cannot be emptied and contribute to ocean water volume changes and resulting sea-level rises), and those residing above sea-level that potentially can be filled with or emptied of groundwater. With respect to the latter, “… only the aquifers are able to absorb, store, and transmit water through their pore spaces and thus participate in the process …” (Hay and Leslie, 1990, p. 166) of climate induced imbalances in the ocean-continent water distribution via the hydrologic cycle. Thus, the available volume depends on the respective eustatic sea level and the average continental elevation at the time in question.

For the present day Earth, Hay and Leslie (1990) gave a value of about $25 \times 10^6$ km$^3$ of pore space within the upper 1 km of average elevation of the continents. This pore space equals (if it could be alternately filled with or emptied of water completely) a global sea-level change of 76 m, or 50 m after applying isostatic adjustment (Hay and Leslie, 1990). It is, thus, approximately equivalent to the total volume of water currently stored in ice sheets, ice caps, and glaciers on land today, though only a proportion of a corresponding water volume is considered to effectively result in sea-level changes; this proportion, however, is significant (see Fig. 2 and below, and Wendler and Wendler, 2016-in this volume, 2016-in this volume; Wendler et al., 2016-in this volume; Wagreich et al., 2014). Operative timescales of aquifer-eustasy are estimated to be $10^4$ to $10^5$ years or <0.01 million years (Hay and Leslie, 1990; Cloetingh and Haq, 2015). This means that amplitudes and operative timescales, and thus rates, of aquifer-eustatic sea-level changes lie within a similar order of magnitude as those for glacio-eustasy (cf. Figs. 2, 3, 4, 5). Hay and Leslie (1990) also expanded on their thoughts by providing hypothetical models for times in the geologic past, including the mid-Cretaceous. These models, based on conservative estimates, suggest that the available pore water volume and retention capacity of aquifers at 200 m average elevation above sea-level could have been twice that of today (see Section 3.1.).

The hypothesis of groundwater-driven eustasy or aquifer-eustasy and its potential to explain short-term eustatic sea-level changes in mid-Cretaceous-like ice free worlds, has been widely disregarded previously because of the underestimation of the water capacity of groundwater aquifer reservoirs on the one hand, and its confusion with the minor and nearly negligible lake and river water volume ($0.03–0.3 \times 10^6$ km$^3$; see Fig. 2) with respect to its sea-level change equivalent (≪1 m) on the other hand (see Hay and Leslie, 1990; Miller et al., 2005a,2005b; Wendler et al., 2016-in this volume). A further reason is that to this day, the processes and efficacy behind climatically controlled groundwater-forced sea-level changes are not well understood, particularly as to their timescales.

However, our understanding of the subject is continuously growing with considerable progress in recent years: Since water content and capacity of the global atmosphere (~25 mm eustatic sea level equivalent, Fig. 2) are thermodynamically constrained, the gain or loss of water by the continents corresponds to an equal loss or gain of water by the oceans (Milly et al., 2010). Excluding continental ice sheets (see Section 2.5.) and anthropogenic causes (cf. Milly et al., 2010), this continent–ocean water exchange is a dynamic process being (more or less) in relative balance, i.e. there is constant backflow of groundwater into the oceans and the aquifers are continuously refilled (Wendler and Wendler, 2016-in this volume). Thus, the process of aquifer-eustasy is based on a dynamic balance between charge (through precipitation) and discharge (through fluvial runoff) of surface and subsurface aquifers that reflect the intensity of the hydrologic cycle (Wendler and Wendler, 2016-in this volume). Consequently, groundwater-driven eustasy or aquifer-eustasy must be driven by imbalances in the ocean–continent water distribution and the hydrologic cycle which, in turn, are climatically controlled. Aquifer-eustasy is, essentially, considered to have been a pervasive process throughout Earth history (Jacobs and Sahagian, 1995; Wendler and Wendler, 2016-in this volume). While both aquifer-eustatic and glacio-eustatic forcing have formed a combined sea-level response during Earth history, aquifer-eustasy outpaces glacio-eustasy during greenhouse phases while remaining active but subsidiary effective during icehouse phases (Wendler and Wendler, 2016-in this volume).

Increases in groundwater storage and corresponding significant short-term aquifer-eustatic sea-level falls occur if the filling processes exceed the draining (aquifer charge > discharge) processes on a global scale of consideration (including associated lake-level rise trends), and the other way around for the emptying of the reservoirs. Acceleration of the hydrologic cycle in particular has been suggested as driving mechanism for sea-level falls caused by longer-term groundwater storage on the continents (e.g. Jacobs and Sahagian, 1993; Föllmi, 2012; Wendler et al., 2011; Wagreich et al., 2014; Wendler and Wendler, 2016-in this volume; Wendler et al., 2016-in this volume), particularly during warm greenhouse climate modes that had little or no ice, such as the mid- to Late Cretaceous (Albian–Santonian, Wendler and Wendler, 2016-in this volume).

Net charge of continental reservoirs, and corresponding eustatic sea-level falls, may thus happen during times of an accelerated hydrological cycle transporting more water towards the continents including the ice-free high latitude areas (Wendler and Wendler, 2016-in this volume). Significant short-term aquifer-eustatic sea-level rises would then be linked to periods of dryer climates and precipitation decrease, when aquifer draining processes exceed the filling processes (aquifer discharge > charge). Wendler et al. (2016-in this volume)) provide the first empirical evidence for a correlation between changes in precipitation, continental weathering intensity, evaporation and astronomically (long-obliquity) forced sea-level cycles during the Cretaceous “Supergreenhouse” (Cenomanian–Turonian) period, making aquifer-eustasy a plausible explanation for short-term eustatic sea-level fluctuations. Nevertheless, many processes behind aquifer-eustasy or other alternatives to glacio-eustasy remain insufficiently understood to date, especially regarding their full complexity and timescales (e.g. considering isostatic rebound effects of the lithosphere through groundwater unloading at the continental margins, see Section 2.7.), and the deceleration of the aquifer discharge.

Additionally, we are largely unable to reconstruct groundwater tables and groundwater-table changes directly from the sedimentary record. Response times of the (constantly flowing) hydrological system to climate changes are short, and can be considered quasi instantaneous given geological timescales and temporal resolution in deep-time. The time interval necessary to fill or empty the continental water reservoirs by an amount equivalent to significant changes in global sea-water volumes, however, may be considerably longer due to complex feedback mechanisms (tens of thousand to hundreds of thousands of years, Hay and Leslie, 1990; cf. Fig. 3 herein). Consequently, Wagreich et al. (2014) indicate a possible lag between a (climate induced) step-function change in the global hydrological cycle and the resulting sea-level changes caused by groundwater storage on land or inflow into the sea. Combining these facts with the obvious conclusion that there should be a positive correlation between filled aquifers (and high groundwater tables) and relatively high lake levels (at least generally on regional to global scales), Wagreich et al. (2014) suggested that non-marine sequences (i.e. lake-level changes as documented in the geologic record) should lie within the longer Milankovitch band (3rd-order cycles), but out-of-phase with sea-level changes. This means that respective lake-level changes record astronomically forced, cyclic climate changes, and should be (mainly?) driven by aquifer-eustasy and thereby record significant groundwater-table changes. This, in turn, would allow for high-resolution, cyclostratigraphic correlation with marine sequences, provided that the non-marine sequences can be sufficiently dated geochronologically. Preliminary tests seem to support this hypothesis (see Wagreich et al., 2014, and Section 3.2. for details).

From this we can conclude that lakes provide a proxy to indirectly record aquifer-eustatic cycles since lake deposits are the best archive available documenting (non-marine) climate cyclicities. Thus, lake-level reconstructions give information on significant groundwater-table changes, and corresponding continent–ocean water distribution imbalances (Wagreich et al., 2014). This led Wagreich et al. (2014) to propose the term “limno-eustasy” as an alternative for aquifer-eustasy used by other authors (e.g. Wendler and Wendler, 2016-in this volume, and references therein), the former being a more all-embracing term for the following reasons: Though “limnic” derives from Ancient Greek for lake (“limne”), the limnologic practice since the 1970s is that the term “limnic” (and the fields of work covered by limnologists) has been extended to cover all inland (also “non-marine”) water bodies — whether they are freshwater or saline, permanent and temporary (ephemeral), flowing (lentic) or standing (lotic), surface or underground (e.g. Elster, 1974; Wetzel, 2001), including aquifers. Consequently, the term “limno-eustasy” would have a wider meaning and not only cover the dominant water volume parameter and driver, but also secondary proxies (reconstructions of lake-level changes and associated groundwater-table changes) of climatically induced periodic changes on land that record groundwater-driven eustatic sea-level changes.

### *2.7. Solid-Earth contributions*

Following Section 2.3.A, this section briefly outlines the solid-Earth factors in more detail, particularly as relevant to short timescale sea-level fluctuations in deep-time. For comprehensive recent overviews see Conrad (2013) and Cloetingh and Haq (2015), and references therein. The relevant key terms as given in Fig. 3 are highlighted by italic type.

*Glacial isostatic adjustment (GIA)* strongly influences eurybatic sea-level measures today and in the recent past (Engelhart et al., 2011). Here, both the ice and water load contributions must be considered (Shennan and Horton, 2002, p. 511). GIA comprises two components, since the Earth responds to the removal (or placement) of a load from (on) its surface in two ways: (1) The elastic response (*elastic rebound of lithosphere*) takes place instantaneously (Conrad and Hager, 1997; Mitrovica et al., 2001), e.g. the recent melting of the Greenland ice sheet (which causes ~0.6 mm per year of global sea-level rise via melt-water inflow, Jacob et al., 2012; Harig and Simons, 2012) causes elastic expansion of the rocks beneath Greenland, leading to 10–30 mm/year of crustal uplift near the most rapidly melting areas (e.g. Bevis et al., 2012; Nielsen et al., 2012). (2) The viscous response (*viscous mantle flow*) takes place subsequently over a timescale of thousands of years ($10^3$–$10^5$ years), e.g. Greenland will continue to uplift (slowly) in response to the current mass loss of its ice sheet. These two processes involve different physical mechanisms of rock deformation that operate on different timescales: the elastic deformation results from changes in the interatomic distances and spaces on a short-term, whereas viscous deformation involves the much slower process of atom migration within the rock. Therefore, unless the (elastic) instantaneous uplift occurring along with the melting (or groundwater release/unloading) is specifically invoked, “isostatic rebound” usually implies the viscous deformation component, and is thus regarded as a viscous process occurring over thousands of years that continues after all the ice is melted. Altogether, Earth's elastic response to ice and water unloading, and the subsequent viscous post-deglaciation response (e.g. the ongoing uplift of Scandinavia), leads to mass redistribution, and thus regional vertical movements along the continental margins (Mitrovica et al., 2001; Conrad, 2013; Haq, 2014), but both processes can be regarded as quasi-instantaneous on geological timescales. Therefore, on the million year and longer timescales of deep-time archives, GIA can be neglected, as long as isostatic compensation of added or removed seawater, which reduces eustatic (global) sea-level change to 70% of its uncompensated value, is included within sea-level change estimates. However, isostatic rebound processes become important for understanding eustatic vs. relative sea-level changes for 100s to tens of thousands of years timescales, especially during the Pleistocene (Miller et al., 2011), and can also influence eustatic sea level because they can affect the net volume of the ocean basins (e.g. Mitrovica and Peltier, 1991).

*Changes to the container capacity of the oceans* operate mainly on long timescales ($10^6$–$10^8$ years and longer), and involve solid-Earth processes. Changes to the volume of the global mid-ocean ridge system is one of the main drivers of long-term global sea-level trends and result from changes in both spreading rates and the total length of the ridge system (Pitman, 1978; Müller et al., 2008; Conrad, 2013). These changes affect the mean age, and thus depth, of oceanic crust. Longer ridge systems and increased seafloor spreading rates (*MORB production rates*) raise the average depth of the sea floor and thus elevate eustatic sea level. Today's volume of mid-ocean ridges elevate sea level by about 570 m, but faster spreading during the Cretaceous produced wider ridges that elevated sea level by up to 820 m (Conrad, 2013). This change resulted in a ~250 m drop in sea level in the last ~125 Ma (Müller et al., 2008). However, more rapid sea-level change with amplitudes of ~50 m occurring over timescales of ~20 Ma require spreading rates to globally accelerate or decelerate by ca. 50% (Conrad, 2013) over these time periods, which may not be tectonically possible, at least globally. Fluctuations in spreading rates may thus explain eustatic sea-level change on ~100 Ma timescales (1st-order sea level cycles), but not on ca. 30 Ma or shorter timescales (2nd-order cycles) because significant changes in average spreading rate occur only over timescales of ~100 Ma and furthermore require similar timescales to offset the average depth of the seafloor.

Secondary effects on the container volume include changes in *ocean floor volcanic activity* (primarily, the emplacement of Large Igneous Provinces, LIPs, during the Cretaceous) and time-varying sediment infill into the oceans (Harrison, 1990; Xu et al., 2006; Müller et al., 2008; Conrad, 2013). Eustatic sea-level rise (or fall) results if the rate of emplacement of volcanics or sediments is faster (or slower) than their rate of removal by subduction. Remarkably, changes in marine sediment volume were considered by Suess (1888) as the main process leading to positive eustatic movements, i.e. rising sea-levels and transgressions. However, large uncertainties are connected to estimates of sediment thickness and the time-dependence of carbonate production and carbonate compensation depth in time (Conrad, 2013). Nevertheless, both Müller et al. (2008) and Conrad (2013) suggested that the net aging of the seafloor since the Cretaceous should have allowed sediments to accumulate, possibly raising sea level by ~60 m. The contribution of seafloor volcanism may have a similar magnitude, but possibly a different time history, raising sea level by up to 100 m during the Cretaceous (Müller et al., 2008) as the Cretaceous LIPs were emplaced on the seafloor, and dropping sea level by ~40 m during the Cenozoic, as the seafloor LIPs are lost to subduction (Conrad, 2013).

Supercontinent cycles are associated with changes in the area of the ocean basins, and may influence sea-level on long-term (>100 Ma) timescales (Conrad, 2013), thus providing the background for first-order sea-level changes. Up to 30 m of sea-level rise may result from the break-up of Pangaea (Kirschner et al., 2010), and a similar drop during the Cenozoic associated with Alpine – Himalayan orogeny (Harrison, 1990). Supercontinent assembly by continental collision and associated orogeny lowers sea level principally by expanding the ocean basin area, resulting in a sea-level drop during assembly and rise during supercontinent dispersal, such as observed during Jurassic and Cretaceous times during Pangaea break-up.

*Mantle flow* supports significant long-wavelength (thousands of kilometers) topographic relief on Earth's surface, with elevated topography occurring above mantle upwelling and depressed topography above downwellings (Hager et al., 1985). Locally, the dynamic submergence or uplift of a coastline results in regional-scale transgressions or regressions (Flament et al., 2013). Globally, this *dynamic topography* also deflects the seafloor in a net sense, and thus offsets sea level. Currently, this offset is positive with an amplitude of up to ~100 m (Conrad and Husson, 2009), because mantle upwellings occur preferentially beneath the seafloor. This sea-level offset may change with time as convection patterns evolve within the mantle, and as the continents migrate. The resulting change in the container volume of the ocean basins results in eustatic sea level change, and rates of up to ~0.5 m/Ma of eustatic sea-level rise have been reconstructed for the past >100 Ma (Conrad, 2013; Conrad and Husson, 2009; Spasojevic and Gurnis, 2012). Although mainly of first or second order duration, dynamic topography may overlap also with 3rd-order sea-level changes regionally, i.e. in the range of a few million years along specific coastlines (e.g. Lovell, 2010).

*Lithosperic flexure and intraplate deformation* involve regional vertical motions and thus affect regional sea level in regions where these processes are important. Spatial and temporal variations in vertical motions in continental interiors as well as along their margins can be modified due to rifting processes, inherited lithospheric structure, and plume emplacement, and thus influence regional sea-levels (Cloetingh and Haq, 2015). Intraplate stresses influence the long-term surface response to mantle upwelling or basal tractions associated with lateral mantle flow, and may also result in short-term, 3rd-order regional changes that overlap with longer-term climate cycles in the few million year range (Cloetingh et al., 1985).

### 2.8. Geoid contributions

In principle, variations in the geoid (an arbitrary gravitational equipotential surface) do not produce a net eustatic sea-level effect. However, local measurements of sea level relative to the continents may be influenced by changes to the geoid. Such changes may result from mass exchanges between the cryosphere and the oceans ("ocean geoid"), which can decrease local gravitational potential near regions of mass loss (e.g. Engelhart et al., 2011), or can perturb Earth's rotation (Milne and Mitrovica, 1998). The "continent geoid" regionally varies through mass exchanges caused by erosion or net charge or discharge of groundwater aquifers. These effects must be considered when accounting for mass movements that cause sea level change, and their associated glacial isostatic adjustments.

### 2.9. Reconstructing sea-level changes in the geologic record

Sea-level changes per se are not recorded unequivocally in the deep-time geological record. In principle, physical, chemical or biological evidence and/or proxies can be used to decipher fluctuations in past sea level. Originally, when defining the term "eustatic", Suess (1888) relied on physical evidence for raised beaches above the prevailing sea-level and shifting fossil shorelines, i.e. fully marine sediments overlying non-marine sediments (Wagreich et al., 2014; Şengör, 2015). Such physical evidence has since been incorporated into the development of sequence stratigraphy, where the reconstruction of shifting shorelines (shoreline trajectories, e.g. Catuneanu et al., 2011) and geometrical evidence for falling and rising sea-levels, and unconformities in coastal sections as an expression of sequence boundaries, provide the building blocks for conceptual and generic types of stratigraphy, especially as used with seismic sections within petroleum industry (e.g. Simmons, 2011, 2012).

In that respect, epicontinental marine basins and flooded continental margins and interiors provide a special setting during greenhouse, high sea-level episodes of Earth history. Especially for the mid- and Late Cretaceous, the number and extent of epicontinental seas was exceptionally high (e.g., Hay and Floegel, 2012). Such basins, like the Western Interior Seaway or the Chalk sea of northwestern Europe, are strongly shaped by complex vertical tectonic movements, which significantly amplify or attenuate the effects of eustatic forcing (Haq and Al-Qahtani, 2005; Zorina, 2014). These shelf seas are characterized by the absence of a continental slope – a key geologic element of oceanic basins – which favors the formation of offlap and onlap stacking patterns. Instead, even minor sea-level changes cause a shoreline to migrate over extremely large distances, resulting in wide (hundreds of kilometers) facies successions, i.e. platformal sequences (Zorina, 2014). Retrogradational parasequence sets may accumulate in basins deepening during regressions, and in those shoaling during transgressions. Consequently, as the architecture of subsequent sequences depends on a complex combination of deepening-shoaling and transgressive–regressive cyclicity, the construction of regional sea-level curves requires a comprehensive analysis of basin evolution including analysis of spatiotemporal facies distribution and reliable estimates of paleo-water depths.

Short term sea-level fall records in carbonate platforms may coincide with longer-term events, with parasequence boundaries superimposed on sequence boundaries. Therefore, separation of short-term and long-term sea-level falls on carbonate platforms is a critical issue, and needs detailed studies of the sedimentary structures (Yilmaz and Altiner, 2001, 2006; Catuneanu et al., 2011; Moore and Wade, 2013). Even the short-term sea-level changes can be affected by regional tectonics. Therefore, some sequence boundaries may not be well preserved over a longer distance. Seismic expressions or geometrical correlations in association with event beds can be more helpful for long distance correlations and for understanding the presence of diachronism related to sequence boundaries.

Apart from the physical evidence provided by stratal geometries and unconformities, paleo-water depths cannot be measured directly in the sedimentary archive (e.g. Burton et al., 1987) except for rare cases of single sedimentary structures like wave ripples. However, facies and facies changes can be related to estimates of depositional water depths in the marine realm, with more confidence and smaller error bars of 2–10 m in the shallow-marine realm (i.e. in the neritic realm, from beach to offshore), and larger errors of tens or hundreds of meters for deep-water environments (bathyal to abyssal). Facies zonations, e.g. in carbonate platforms with reefs, lagoons and fore-reef facies, and evidence for supra-, intra- and subtidal deposition and photic zone carbonate production can be helpful.

Paleontology and micropaleontology, given primary taphocoenoses, may provide further evidence by presenting depth-restricted biota and assemblages (e.g. sea grass and associated faunas, Hart et al., 2016-in this volume). Foraminiferal assemblages may provide relatively precise indicators for depositional water depths, and are especially useful in deeper-water sediments (e.g. Murray, 1991; Sliter and Baker, 1972; Hart, 1980; Koutsoukos and Hart, 1990; Widmark and Speijer, 1997; Abramovich et al., 2003; Kaminski and Gradstein, 2005). However, reconstructions of sea-level changes in the pelagic to hemipelagic realm, at bathyal water depths below 150–200 m, are considerably hampered by the fact that (1) depositional water depths (several 100 s to 1000 m) largely exceed the magnitude of inferred sea-level changes, (2) correlative conformities mark sequence boundaries in bathyal environments instead of unconformities that are easily recognizable in coastal areas and carbonate platforms, thus, changes in sedimentation may be subtle and not discernable by lithofacies, and (3) although present, trends in fossil communities related to changes in depositional depths may not be as obvious and clear as in coastal areas, and may become more and more subtle and harder to recognize (e.g. Wolfgring et al., 2016-in this volume).

Beyond that, chemical and mineralogical proxies are increasingly used in fine-grained shelf to bathyal sediments to decipher sea-level changes. Along the shelf-slope-basin profile, climate as well as carbonate versus siliciclastic domination of the system has to be taken into account when using chemical proxies for interpreting sea-level changes. In principle, times of sea-level lowstands may be characterized by sediments with generally higher siliciclastic contents, coarser grain sizes of siliciclastics and higher clay contents. Transgression results in condensed sections and may occur in low oxygen environments. Thus, high terrigenous clay mineral peaks may record sea level lows. Various chemical proxies include carbonate content, Sr/Ca ratio (Li et al., 2000), uranium content, carbon and oxygen isotopes, Si/Al, Ti/Al, Zr/Al, Zr/Ti, Mn and Mn/Al ratios (see Jarvis et al., 2001; Olde et al., 2015).

### *2.10. Constructing short-term sea-level curves from the geologic record*

Short-term eurybatic sea-level reconstructions and sea-level shift amplitudes are based mainly on sequence-stratigraphic data from around the world, including outcrops, well-logs and seismic profiles (see Haq, 2014 and references therein for details; also e.g. Haq et al., 1987; Hardenbol et al., 1998; Simmons, 2011, 2012). Correlations of regional sea-level curves, reinforced by oxygen-isotopic trends, provide means of recognizing synchronous global sea-level events (e.g. Haq, 2014). In addition, sea-level sensitive facies and seismic geometries are used to identify sea-level changes, i.e. condensed section deposits such as organic-rich sediments, transgressive coals, evaporites, carbonate megabreccias, exposure-related deposits such as karst and laterite, forced regressive facies, and radiations, extinctions or migrations of shallow marine faunas are used in reconstructions (Haq, 2014). Amplitudes of eustatic sea-level changes are estimated based on averages of eurybatic measurements for rises and falls from all stratigraphic sections under consideration (op. cit.). As these measurements are always imprecise, Haq and Schutter (2008) classified each event quasi-quantitatively by measuring the amount of fall from the previous highstand, and classified respective events as minor (<25 m), medium (25 to 75 m), or major (>75 m), and Haq (2014) adopted this scheme for his revision of Cretaceous eustasy and the revised Cretaceous 3rd-order sea-level curve.

## 3. The Cretaceous world

The Cretaceous period represents the youngest prolonged greenhouse interval in Earth history (e.g., Skelton, 2003; Hay, 2008). Greenhouse climate is attributed to elevated $CO_2$ (and other greenhouse gases) levels, with 2–16 times the pre-industrial level (Hay and Floegel, 2012). Pole to equator temperature gradients were reduced, with mostly relatively warm polar regions. Long-term sea-level was high, about 170–250 m above present sea-level (Conrad, 2013; Haq, 2014), mainly a result of rapid spreading rates at mid-ocean ridges.

Paleoceanographic and paleogeographic changes accompanied the final breakup of Gondwana during the Cretaceous, and the opening of the South Atlantic and the Indian oceans, and other complications related to the opening and closing of Tethyan basins. The latter provides a major oceanic gateway for circulation, connecting the mid-latitudinal Atlantic to the Caribbean and the Pacific. Hadley cell shrinkage (Hasegawa et al., 2012; Hay and Floegel, 2012), restricted thermohaline circulation (e.g. Friedrich et al., 2008), and the possible presence of oceanic eddies (Hay, 2008) may have resulted in a climate-ocean system very different from today's (Hay and Floegel, 2012). The paleogeographic situation was characterized by flooded continents, large and shallow epicontinental seas, and large marine seaways. For the later part of the Late Cretaceous, the opening of the South Atlantic for deepwater circulation changed the paleoceanographic pattern considerably (Friedrich et al., 2012).

Recent research indicates the presence of 3–4 climate states (Kidder and Worsley, 2010, 2012; Hu et al., 2012; Hay and Floegel, 2012), i.e. a cooler greenhouse during the early Early Cretaceous (Berriasian–Barremian), a very warm greenhouse in the mid-Cretaceous (Aptian–Turonian/Coniacian) including short-lived hothouse periods with widespread anoxia (OAE1, 2 and 3) and possible reversals of the thermohaline circulation (HEATT episodes of haline euxinic acidic thermal transgression, Kidder and Worsley, 2010), and a (later) Late Cretaceous (Santonian/Campanian–Maastrichtian) evolution from warm to cool greenhouse. In addition, an increasing number of short-term climatic events within the longer term trends are reported for the Cretaceous (e.g. Hu et al., 2012).

Cretaceous sea-level changes have been investigated more recently by Cloetingh and Haq (2015), Haq (2014), Immenhauser (2005), Miller et al. (2005a, b, 2009), Kominz et al. (2008), and Müller et al (2008), but the (global) correlation and significance of these sea-level changes are still arguable (e.g., Zorina et al., 2008; Lovell, 2010; Petersen et al., 2010; Boulila et al., 2011; Haq, 2014).

### *3.1. Cretaceous short-term sea-level changes and their drivers*

The timing, the causes, and the consequences of significant short-term (i.e. several thousand to 100 s of ka) sea-level changes during this last major greenhouse episode of Earth history are strongly debated issues. A major episode of oceanic crust production during and after the breakup of Pangaea led to long-term sea-level rise and a highstand during Cretaceous times. Peak sea level during the Cretaceous is estimated between 85 and 280 m, with best estimates between 170 and 250 m, above today's sea level (Müller et al., 2008; Miller et al., 2011; Conrad, 2013; Haq, 2014). Our current state of knowledge is that solid-Earth dynamics that are not related to glacio- or hydro-isostasy (Section 2.7.) can well explain first-order sea-level cycles, and probably contribute to second order cycles (we still do not have a good explanation for 2nd-order cycles, see Conrad, 2013), but cannot explain the prevalent 3rd-order cycles evident from, e.g. Cretaceous (Haq, 2014), sequence stratigraphy. However, short-term, 3rd- to 4th-order sea-level changes, recorded in Cretaceous strata, could exhibit amplitudes similar to those of Pleistocene glacial–interglacial episodes, i.e. 15–50 m (Miller et al., 2005b; Kominz et al., 2008), and qualify as minor to medium according to Haq and Schutter (2008; see also Section 2.10).

Although debate regarding the existence of Cretaceous eustatic (globally synchronous) sea-level change persists (e.g. Moucha et al., 2008; Lovell, 2010; Ruban et al., 2010; Zorina et al., 2008), Haq (2014) states that eustasy cannot be dismissed in the Cretaceous. This is based on the growing evidence that at least some if not all 3rd-order sequences, even during the extreme hothouse episode ("Supergreenhouse") of the mid-Cretaceous (e.g. Hay and Floegel, 2012), were synchronous (see most recent compilation by Haq, 2014; Wilmsen and Nagm, 2013; Wendler et al., 2014), and therefore record short-term eustatic sea-level changes. As discussed above (Sections 2.5. and 2.6.), two hypotheses may explain the major processes controlling such eustatic sea-level changes: glacio-eustasy and aquifer eustasy. Additional mechanisms that should be considered are thermo-eustasy (both thermosteric and halosteric effects surely played a role during extreme warm of the Cretaceous, but may be confined to a few meters of change), and sediment input and LIPs emplacement. Sediment input in particular can also act on short-term timescales, but on such short timescales its impact is likely limited to a maximum of a few meters (Conrad, 2013).

For the Early Cretaceous with its generally cool greenhouse climate (Hay and Floegel, 2012; Hu et al., 2012; Föllmi, 2012) glacio-eustasy seems to be a likely driver for short-term sea-level changes (cf. Section 2.5.) given the presence of direct evidence for ice (e.g. Alley and Frakes, 2003) and indirect evidence for cool (marine) temperatures such as glendonites (Price and Nunn, 2010), stable oxygen isotope data (Stoll and Schrag, 1996) and Cretaceous Oceanic Red Beds (Wagreich, 2009). The same may hold true for the later part of the Late Cretaceous, the Campanian–Maastrichtian, when especially sequence stratigraphy and correlated stable oxygen isotopes indicate cool intervals with sea-level lowstands (Miller et al., 2005b; Bowman et al., 2013). Thus, the Early Cretaceous and possibly also the later part of the Late Cretaceous can be identified as times of a cool greenhouse climate with the possibility of ephemeral ice sheets on Antarctica and maybe also on parts of Siberia (Miller et al., 2005b; Hay and Floegel, 2012).

For the mid-Cretaceous, especially the hottest period of the Mesozoic during the Cenomanian–Turonian, a warm (Super-)greenhouse state with common hothouse intervals connected with oceanic anoxic events is reconstructed (Hay and Floegel, 2012). Both for the Cenomanian (Moriya et al., 2007) and for the Turonian (MacLeod et al., 2013), continuous stable oxygen isotope records from excellently preserved glassy foraminifera do not show any inferred ice-induced oxygen isotope shifts and strongly argue against the presence of even ephemeral ice sheets. Thus, at least for the Cenomanian–Turonian, alternative processes like aquifer-eustasy have been invoked to explain short-term sea-level changes (Wagreich et al., 2014; Wendler et al., 2014; Wendler and Wendler, 2016-in this volume; Wendler et al., 2016-in this volume).

Amplitudes, operative timescales, and rates of (mid-)Cretaceous aquifer-eustatic sea-level changes may have been significantly larger as based on conservative estimates by Hay and Leslie (1990), as the available pore water volume and retention capacity of aquifers at 200 m average elevation above sea-level could have been twice that of today (~$40 \times 10^6$ km$^3$). This would double the resulting maximum amplitudes of corresponding eustatic sea-level changes (up to about 80 m), even though for the Cretaceous, such storage estimates exclude aquifers below 200 m continental elevation as being unavailable for groundwater charge and discharge because of the higher sea-level and associated permanent saturation with water (Conrad, 2013; Wendler et al., 2016-in this volume). In addition, it can be assumed that during the Cretaceous greenhouse the expanse of deserts was smaller and ice-free polar regions were additionally available for aquifer charge and discharge on the one hand, and due to higher global temperatures an enhanced hydrological cycle transported more water towards these high latitudes (e.g. Flögel et al., 2011; Suarez et al., 2011; Wendler et al., 2016-in this volume) on the other hand. Altogether, there was tremendous potential for continental water storage during the Cretaceous warm greenhouse.

Short-term sea-level cycles in the longer Milankovitch band are increasingly being recognized in the Mesozoic and Cenozoic deep-time records (e.g. Boulila et al., 2011), thus requiring climatically controlled drivers for sea-level (and arguing against purely regional sea-level fluctuations). Fourth-order cyclicity seems to be related mainly to the 405 ka periodicity, which most likely represents long-period orbital eccentricity control on sea level and depositional cycles. Third-order cyclicity, expressed as time-synchronous sea level falls of ~20 to 110 m on ~0.5–3.0 Ma timescales in the Cretaceous (Haq, 2014) could be related to climate cycles on the long Milankovitch scale, i.e. 1.2 and 2.4 Ma orbital cycles (e.g. Boulila et al., 2011; Wendler et al., 2014; Olde et al., 2015). Longer-term cycles, e.g. a 4.7 Ma band, are also archived in the carbon isotope records and may have influenced sea level (e.g. Sprovieri et al., 2013).

### 3.2. Cretaceous short-term eustatic changes as a stratigraphic tool

Regional and global, short-term and long-term sea-level changes are displayed as sedimentary sequences in the geologic record, the character of which can be cyclic (i.e. having a certain frequency) or non-cyclic. Nonetheless, it is important to clarify that “sequence stratigraphy and eustasy are separate (although related) concepts” (Simmons, 2012, p. 240). The principles of sequence stratigraphy can, notwithstanding, be regionally (intra-basinal) applied without reference to driving mechanisms (Simmons, 2012). Sequence stratigraphy is based on attempts to subdivide sedimentary successions into packages relating to changes in (eurybatic) sea-level at a variety of scales, while eustasy describes an understanding of globally synchronous sea-level change (op. cit.).

For supraregional correlations (i.e. inter-basinal to global), we need chronostratigraphic and/or geochonologic tools to correlate the sequences, that is to say, the sequences need to be rooted in time-stratigraphy for correlation (see Simmons, 2012). The signal of short-term eustatic sea-level change is theoretically well-suited for that purpose because it is global and synchronous. However, as this signal is cyclic, additional tools are needed to date and correlate respective repeating sequences, i.e. to provide a chronostratigraphic and geochronologic framework of sufficient resolution. Moreover, as elaborated in Section 2, the eurybatic sea-level change signal in the geologic record results from a complex combination of processes that can cancel out/decelerate or amplify/accelerate each other to produce an underlying eustatic signal. To differentiate eustatic from eurybatic signals is the main challenge in supraregional sequence stratigraphy and cyclostratigraphy.

A primary application, and evidently connected to the sea-level reconstructions, is the wide usage of sequence stratigraphy in the petroleum industry (see Simmons, 2011, 2012). Here, primarily, sequence stratigraphy provides a tool for regional, mostly intra-basinal correlations and predictions combining various datasets within an integrated framework (Simmons, 2012). However, from the huge amount of regional datasets available from the petroleum industry, a global sequence stratigraphic framework emerged out of regional studies around the world that led to the recognition of eustasy in the geologic record on the one hand, and a eustatic sea-level curve on the other hand (Haq et al., 1987; Haq, 2014; Hardenbol et al., 1998). Although debatable in its details (e.g. Simmons, 2012 and references therein), this stresses the importance of synchronicity of processes in the geologic-stratigraphic record. In proving synchronicity, a global sequence stratigraphic scheme and sea-level curve becomes a basis for chronostratigraphic correlation itself, especially if connected to other globally applicable stratigraphic methods like biostratigraphy, chemostratigraphy (stable carbon and oxygen isotope stratigraphy, e.g. Saltzman and Thomas, 2012; Grossman, 2012) or magnetostratigraphy (e.g. Ogg, 2012).

It follows that time-rooted sequence stratigraphy is the link between eustasy and stratigraphy. On this basis, sequence stratigraphy is naturally connected to cyclostratigraphy, i.e. the study and application of astronomically forced cycles such as widely recognized ‘Milankovitch cycles’ (orbital timescales, astrochronology, e.g. Hilgen et al., 2015; Hinnov, 2013; Hinnov and Hilgen, 2012) of global climate and ocean circulation patterns that are displayed in sedimentary successions (sequences).

Astrochronology, as based on (climate) cyclostratigraphy, is one of the major stratigraphic tools for establishing a stable and high-precision geological time scale (e.g. GTS12, Gradstein et al., 2012; see also projects like EARTHTIME and Kuiper et al., 2008; Laskar et al., 2011; Waltham, 2015). In this respect, sea-level cycles that are related to Milankovitch-type orbital cycles, provide the means for absolute dating in the stratigraphic record, especially in the 100 ka and longer Milankovitch band cycle frequencies that have been demonstrated for Cretaceous sequences (e.g. Boulila et al., 2011; Wendler et al., 2014).

Short-term climate cyclicity during the Cretaceous is recorded by cyclic sedimentation such as limestone–marl cycles with periods in the Milankovitch bands of mainly precession (ca. 20 ka) and eccentricity (100 and 405 ka) and longer such as 1.2 Ma and ca. 2.0 Ma. For the later part of the Late Cretaceous, from the Cretaceous–Paleogene boundary backwards, cyclostratigraphic records and astrochronology are well established (e.g. Hennebert et al., 2009; Voigt and Schönfeld, 2010; Batenburg et al., 2014; Husson et al., 2011; Wagreich et al., 2012). Backwards in the stratigraphic record, floating orbital time scales of various bits and pieces do exist (e.g. Martinez et al., 2013, 2015; Locklair and Sageman, 2008; Sageman et al., 2006, 2014; Wissler et al., 2004).

### 3.3. Cretaceous cyclostratigraphy and marine to non-marine correlations

Progress in Cretaceous climate change and marine cyclostratigraphy (see above) as well as progress in non-marine (bio-)stratigraphy (e.g. Sames and Horne, 2012) has led to changing concepts, approaches and new hypotheses for proxies and methods for improved marine to non-marine correlations. In principle, these are based on the single synchronous, continuous signal recorded by various proxies in both marine and non-marine successions: astronomically forced, cyclic, short-term (<1 Ma) and medium-term (a few Ma) global climate change. Among others, such methods include the analysis of lake-level fluctuations that are considered to have an out-of-phase interrelationship with short-term sea-level fluctuations during warm greenhouse climate ("limno-eustasy" of Wagreich et al., 2014).

Another approach is non-marine "ecostratigraphy" in interdisciplinary approaches aimed at a non-marine cyclostratigraphy, with geochronologic and magnetostratigraphic control (e.g. Sames, 2015). The consideration that paleoenvironmental changes – which control assemblage changes of microfossils along with changes of lithological and geochemical parameters of corresponding sedimentary successions – are climatically and thus, ultimately, astronomically controlled, leads to the coherent approach that changes can be analyzed for cyclicity and tested for cyclostratigraphic use. Due to the general ephemerality (on geologic timescales) and characteristic strong lateral facies change of non-marine deposits, analyses for cyclicities must be based on multiple proxies (e.g. Sames, 2015; research in progress).

Altogether, non-marine Cretaceous astrochronology is still in its infancy as to the number of studies available, and time intervals covered. Few studies do exist, mostly from long-term lake deposits (e.g. Wu et al., 2013). However, the relevant data basis in the non-marine realm is constantly improving due to ongoing projects, such as the International Continental Drilling Programme (ICDP) Project in the Songliao Basin (NE China), the "Songliao Basin Drilling Project" (e.g. Wang et al., 2009). A time-calibrated non-marine Cretaceous cyclostratigraphy is considered an important tool for high-resolution marine to non-marine correlations in the near future, as well as an important contribution for unravelling short-term Cretaceous climate and sea-level change, e.g. further testing of the aquifer/limno-eustatic hypothesis (Wagreich et al., 2014; Wendler and Wendler, 2016-in this volume).

## 4. Conclusion and perspectives

Various regional and global processes influence sea level, which is the critical interface between three of Earth's main domains, the hydrosphere, the geosphere, and the atmosphere, and also a crucial zone of the biosphere. Sea level is also a critical interface with respect to its relevance for mankind. UNESCO IGCP 609 centers on the fossil greenhouse record of fluctuations in that interface, expressed in sedimentary cycles (sequences) governed by short-term (<0.5 to 3 Ma) eustatic (i.e. global) sea-level changes of the Cretaceous. The Cretaceous, as the last prolonged greenhouse episode of Earth history, contains evidence for significant short-term eustatic sea-level fluctuations that follow Milankovitch cycles, i.e. in the fourth order (mainly 405 ka) and third order (mainly 1.2, 2.4 Ma) range. Provided chronological linking, these cyclic climate (and sea- and lake-level) fluctuations play an important role for high-resolution Cretaceous marine chronostratigraphy with considerable potential for marine to non-marine correlations.

Although continental ice may be the main driver for some of the short-term sea-level shifts during the early Early Cretaceous and the late Late Cretaceous with cool greenhouse conditions, the presence of large continental ice shields is highly unlikely for the warm greenhouse to hothouse conditions during the mid-Cretaceous. Alternatively to glacio-eustasy, aquifer-eustasy may have played a significant role during Cretaceous warm greenhouse and hothouse times, by storing water as groundwater (and lakes) on the continents. This alternative mechanism must be tested in the stratigraphic record, i.e. by relating lake (groundwater) levels to sea level or by applying methods to identify the predominance of humid versus arid climates, i.e. by reconstructing continental weathering related to sequence stratigraphy. In this regard, marine to non-marine stratigraphic correlations with high resolution and precision, and based on Milankovitch climate cycles, have become an essential tool and a prerequisite for evaluating the aquifer-eustasic hypothesis.

Identifying additional processes that influence sea-level changes, especially those effective during greenhouse climate phases of the Earth System, and possibly contributing to recent sea-level rise due to atmospheric greenhouse gas accumulation and associated global warming, is a primary concern for society. To predict future sea-levels in the Anthropocene, we need a better understanding of the record of past sea-level change, especially considering a shift from icehouse to greenhouse climate conditions. Calculations of rates of sea-level change during the Cretaceous greenhouse episode are challenging, but these rates of geologically short-term sea-level change on a warm Earth will help to better evaluate recent global change and, further, to assess the role of feedback mechanisms on sea level.

Water cycling between the oceans and continental aquifers, which may have exerted an important control (aquifer-eustasy) on mid-Cretaceous sea level, may also be important for modern and future, i.e. Anthropocene, sea-level change. For example, groundwater depletion during the past two decades contributed ~0.4–0.6 mm/a to global sea level rise (Konikow, 2011; Wada et al., 2012). Such rates of groundwater transfer to the oceans are comparable to rates of water transfer from the cryosphere (Slangen et al., 2014), and induce an elastic deflection of the solid Earth than can be detected geodetically (Jensen et al., 2013). During the past century, groundwater depletion into the oceans was partially offset by water impoundment within artificial reservoirs (Chao et al., 2008), which also led to significant solid-Earth deformations (Fiedler and Conrad, 2010). However, in the past few decades accelerating groundwater depletion has overwhelmed a slowing rate of water impoundment (Pokhrel et al., 2012), resulting in rates of continental water loss that could approach 1 mm/yr in the coming century (Wada et al., 2012). If perpetuated over millennia, such rates could eventually raise sea level by several meters. Thus far, groundwater depletion is thought to be primarily human-induced by groundwater pumping (Wada et al., 2010). However, given the possibility that the periodic drainage of continental aquifers into the oceans may have been an important aspect of the mid-Cretaceous greenhouse, it is possible that the same sort of aquifer-induced sea-level variations may be important during future greenhouse conditions. Such aquifer-eustatic contributions to sea level could add to the thermo-eustasic and glacio-eustasic contributions that are already expected for a warmer future climate, and elevate projected future sea level beyond current expectations. Such a possibility adds urgency to understanding the mechanisms that governed sea-level change during greenhouse climates in Earth's geologic past, such as during the mid-Cretaceous.

## 5. Author contributions

Idea, concept, research and main writing: BS. Substantiation and second main writing: MW. Main additional contributions: JEW, BUH, CPC; further contributions to various special topics by MCM, XH, IW, EW, IÖY, SOZ.

## Acknowledgements

This paper is a contribution to UNESCO-IUGS IGCP project 609 "Climate-environmental deteriorations during greenhouse phases:

Causes and consequences of short-term Cretaceous sea-level changes" (MW, XH, IÖY, SOZ, BS). It benefited from an ÖAW (Austrian Academy of Sciences) IGCP grant to Michael Wagreich (MW, BS), from Austrian Science Fund (FWF) projects P 27687-N29 (BS) and P 24044-N24 (MW, EW), NSF grant EAR-1151241 (CPC), EARTHTIME-EU workshop funding, and the Russian Government Program of Competitive Growth of Kazan Federal University (SOZ).

## References

Abramovich, S., Keller, G., Stüben, D., Berner, Z., 2003. Characterization of late Campanian and Maastrichtian planktonic foraminiferal depth habitats and vital activities based on stable isotopes. Palaeogeogr. Palaeoclimatol. Palaeoecol. 202 (1–2), 1–29.

Alley, N.F., Frakes, L.A., 2003. First known Cretaceous glaciation: Livingston Tillite Member of the Cadnaowie Formation, South Australia. Aust. J. Earth Sci. 50 (2), 139–144.

Alley, R.B., Anandakrishnan, S., Christianson, K., Horgan, H.J., Muto, A., Parizek, B.R., Pollard, D., Walker, R.T., 2015. Oceanic forcing of ice sheet retreat: West Antarctica. Annu. Rev. Earth Planet. Sci. 43, 7.1–7.25.

Batenburg, S.J., Gale, A.S., Sprovieri, M., Hilgen, F.J., Thibault, N., Boussaha, M., Orue-Etxebarria, X., 2014. An astronomical time scale for the Maastrichtian based on the Zumaia and Sopelana sections (Basque country, northern Spain). J. Geol. Soc. 171, 165–180. http://dx.doi.org/10.1144/jgs2013-015.

Bevis, M., Wahr, J., Khan, S.A., Madsen, F.B., Brown, A., Willis, M., Kendrick, E., Knudsen, P., Box, J.E., van Dam, T., Caccamise II, D.J., Johns, B., Nylen, T., Abbott, R., White, S., Miner, J., Forsberg, R., Zhou, H., Wang, J., Wilson, T., Bromwich, D., Francis, O., 2012. Bedrock displacements in Greenland manifest ice mass variations, climate cycles and climate change. Proc. Natl. Acad. Sci. 109 (30), 11944–11948. http://dx.doi.org/10.1073/pnas.1204664109.

Boulila, S., Galbrun, B., Miller, K.G., Pekar, S.F., Browning, J.V., Laskar, J., Wright, J.D., 2011. On the origin of Cenozoic and Mesozoic "third-order" eustatic sequences. Earth Sci. Rev. 109, 94–112.

Bowman, V.C., Francis, J.E., Riding, J.B., 2013. Late Cretaceous winter sea ice in Antarctica? Geology 41, 1227–1230.

Broeker, W.S., 1991. The great ocean conveyor. Oceanography 4, 79–89.

Burton, R., Kendall, C.G.S.C., Lerche, I., 1987. Out of our depth: on the impossibility of fathoming eustasy from the stratigraphic record. Earth Sci. Rev. 24, 237–277.

Caffrey, M., Beavers, R., 2013. Planning for the impact of sea-level rise on U.S. national parks. Park. Sci. 30 (1), 6–13 (http://www.nature.nps.gov/parkscience/Archive/PDF/Article_PDFs/ParkScience30%281%29Summer2013_6-13_CaffreyBeavers_3647.pdf).

Catuneanu, O., Galloway, W.E., Kendall, C.G.S.C., Miall, A.D., Posamentier, H.W., Strasser, A., Tucker, M.E., 2011. Sequence stratigraphy: methodology and nomenclature. Newsl. Stratigr. 44, 173–245.

Cazenave, A., Le Cozannet, G., 2014. Sea level rise and its coastal impacts. Earth's Future 2, 15–34. http://dx.doi.org/10.1002/2013EF000188.

Cazenave, A., Llovel, W., 2010. Contemporary sea level rise. Ann. Rev. Mar. Sci. 2, 145–173.

Chao, B.F., Wu, Y.H., Li, Y.S., 2008. Impact of artificial reservoir water impoundment on global sea level. Science 320 (5873), 212–214. http://dx.doi.org/10.1126/science.1154580.

Church, J.A., Roemmich, D., Domingues, C.M., Willis, J.K., White, N.J., Gilson, J.E., Stammer, D., Köhl, A., Chambers, D.P., Landerer, F.W., Marotzke, J., Gregory, J.M., Suzuki, T., Cazenave, A., Le Traon, P.-Y., 2010. Ocean temperature and salinity contributions to global and regional sea-level change. In: Church, J.A., Woodworth, P.L., Aarup, T., Wilson, W.S. (Eds.), Understanding Sea-Level Rise and Variability, 1st ed. Wiley-Blackwell, Chichester, pp. 143–176.

Church, J.A., Clark, P.U., Cazenave, A., Gregory, J.M., Jevrejeva, S., Levermann, A., Merrifield, M.A., Milne, G.A., Nerem, R.S., Nunn, P.D., Payne, A.J., Pfeffer, W.T., Stammer, D., Unnikrishnan, A.S., 2013. Chapter 13: Sea Level Change. In: Stocker, T.F., Qin, D., Plattner, G.-K., Tignor, M., Allen, S.K., Boschung, J., Nauels, A., Xia, Y., Bex, V., Midgley, P.M. (Eds.), Climate Change 2013: The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change. Cambridge University Press, Cambridge and New York, pp. 1137–1216.

Cloetingh, S., Haq, B.U., 2015. Inherited landscapes and sea level change. Science 347 (6220). http://dx.doi.org/10.1126/science.1258375 (1258375-10).

Cloetingh, S., McQueen, H., Lambeck, K., 1985. On a tectonic mechanism for regional sea level variations. Earth Planet. Sci. Lett. 75, 157–166. http://dx.doi.org/10.1016/0012-821X(85)90098-6.

Conrad, C.P., 2013. The solid Earth's influence on sea-level. GSA Bull. 125 (7–8), 1027–1052. http://dx.doi.org/10.1130/B30764.1.

Conrad, C.P., Hager, B.H., 1997. Spatial variations in the rate of sea level rise caused by the present-day melting of glaciers and ice sheets. Geophys. Res. Lett. 24 (12), 1503–1506. http://dx.doi.org/10.1029/97gl01338.

Conrad, C.P., Husson, L., 2009. Influence of dynamic topography on sea level and its rate of change. Lithosphere 1 (2), 110–120. http://dx.doi.org/10.1130/L32.1.

Crowley, J.W., Gérault, M., O'Connell, R.J., 2011. On the relative influence of heat and water transport on planetary dynamics. Earth Planet. Sci. Lett. 310 (3–4), 380–388. http://dx.doi.org/10.1016/j.epsl.2011.08.035.

Dewey, J.F., Pitman, W.C., 1997. Sea-level changes: mechanisms, magnitudes and rates. SEPM Spec. Publ. 58, 95–127. http://dx.doi.org/10.2110/pec.98.58.0001.

El Raey, M., Dewidar, Kh., El Hattab, M., 1999. Adaption to the impacts of sea level rise in Egypt. Clim. Res. 12, 117–128. http://dx.doi.org/10.1023/A:1009684210570.

Elster, H.-J., 1974. History of limnology. Mitt. Int. Ver. Theor. Angew. Limnol. (Int. Assoc. Theor. Appl. Limnol.) 20, 7–30.

Emery, K., Aubrey, D., 1991. Sea-Levels, Land Levels and Tide Gauges. Springer-Verlag, New York.

Engelhart, S.E., Horton, B.P., Kemp, A.C., 2011. Holocene sea level changes along the United States' Atlantic Coast. Oceanography 24 (2), 70–79. http://dx.doi.org/10.5670/oceanog.2011.28.

Farrell, W.E., Clark, J.A., 1976. On postglacial sea level. Geophys. J. R. Astron. Soc. 46 (3), 647–667. http://dx.doi.org/10.1111/j.1365-246X.1976.tb01252.x.

Fiedler, J.W., Conrad, C.P., 2010. Spatial variability of sea level rise due to water impoundment behind dams. Geophys. Res. Lett. 37, L12603. http://dx.doi.org/10.1029/2010gl043462.

Flament, N., Gurnis, M., Müller, R.D., 2013. A review of observations and models of dynamic topography. Lithosphere 5, 189–210. http://dx.doi.org/10.1130/L245.1.

Flögel, S., Wallmann, K., Kuhnt, W., 2011. Cool episodes in the Cretaceous — exploring the effects of physical forcings on Antarctic snow accumulation. Earth Planet. Sci. Lett. 307 (3–4), 279–288. http://dx.doi.org/10.1016/j.epsl.2011.04.024.

Föllmi, K., 2012. Early Cretaceous life, climate and anoxia. Cretac. Res. 35, 230–257. http://dx.doi.org/10.1016/j.cretres.2011.12.005.

Friedrich, O., Erbacher, J., Moriya, K., Wilson, P.A., Kuhnert, H., 2008. Warm saline intermediate waters in the Cretaceous tropical Atlantic Ocean. Nat. Geosci. 1, 453–457. http://dx.doi.org/10.1038/ngeo217.

Friedrich, O., Norris, R.D., Erbacher, J., 2012. Evolution of middle to Late Cretaceous oceans — a 55 m.y. record of Earth's temperature and carbon cycle. Geology 40, 107–110.

Gale, A.S., 1996. Turonian correlation and sequence stratigraphy of the Chalk in southern England. Geol. Soc. Lond., Spec. Publ. 103, 177–195.

Gale, A.S., Hardenbol, J., Hathway, B., Kennedy, W.J., Young, J.R., Phansalkar, V., 2002. Global correlation of Cenomanian (Upper Cretaceous) sequences: evidence for Milankovitch control on sea level. Geology 30, 291–294.

Gehrels, W.R., Shennan, I., 2015. Sea level in time and space: revolutions and inconvenient truths. J. Quat. Sci. 30 (2), 131–141. http://dx.doi.org/10.1002/jqs.2771.

Gleick, P.H., 1996. Water resources. In: Schneider, S.H. (Ed.), Encyclopedia of Climate and Weather. Oxford University Press, New York, pp. 817–823.

Gradstein, F.M., Ogg, J.G., Schmitz, M.D., Ogg, G.M., 2012. The Geological Time Scale 2012. Elsevier, Amsterdam (1142 pp.).

Grossman, E.L., 2012. Applying Oxygen Isotope Paleothermometry in Deep Time. In: Ivany, L.C., Huber, B.T. (Eds.), Reconstructing Earth's Deep-Time Climate – The State of the Art in 2012. Paleontological Society Papers 18, pp. 39–67.

Gurnis, M., 1990. Bounds on global dynamic topography from Phanerozoic flooding of continental platforms. Nature 344 (6268), 754–756. http://dx.doi.org/10.1038/344754a0.

Gurnis, M., 1993. Phanerozoic marine inundation of continents driven by dynamic topography above subducting slabs. Nature 364 (6438), 589–593. http://dx.doi.org/10.1016/doi:10.1038/364589a0.

Hager, B.H., Clayton, R.W., Richards, M.A., Comer, R.P., Dziewonski, A.M., 1985. Lower mantle heterogeneity, dynamic topography and the geoid. Nature 313 (6003), 541–545. http://dx.doi.org/10.1038/313541a0 -.

Hallberg, R., Adcroft, A., Dunne, J.P., Krasting, J.P., Stouffer, R.J., 2013. Sensitivity of twenty-first-century global-mean steric sea level rise to ocean model formulation. J. Clim. 26, 2947–2956.

Haq, B.U., 2014. Cretaceous eustasy revisited. Glob. Planet. Chang. 113, 44–58. http://dx.doi.org/10.1016/j.gloplacha.2013.12.007 -.

Haq, B.U., Al-Qahtani, A.M., 2005. Phanerozoic cycles of sea-level change on the Arabian Platform. Geoarabia 10 (2), 127–160.

Haq, B.U., Schutter, S.R., 2008. A chronology of Paleozoic sea-level changes. Science 322, 64–68. http://dx.doi.org/10.1126/science.1161648.

Haq, B.U., Hardenbol, J., Vail, P.R., 1987. Chronology of fluctuating sea levels since the Triassic. Science 235, 1156–1167.

Hardenbol, J., Thierry, J., Farley, M.B., Jacquin, T., Graciansky, P.C., de Vail, P.R., 1998. Mezozoic and Cenozoic sequence chronostratigraphic framework of European basins. SEPM Special Publication 60. Society for Sedimentary Geology, Tulsa, pp. 763–786.

Harig, C., Simons, F.J., 2012. Mapping Greenland's mass loss in space and time. Proc. Natl. Acad. Sci. 109 (49), 19934–19937. http://dx.doi.org/10.1073/pnas.1206785109.

Harrison, C.G.A., 1990. Long-term eustasy and epeirogeny in continents. In: Revelle, R.R. (Ed.), Sea-Level Change. National Academy Press, Washington, D.C., pp. 141–158.

Harrison, C.G.A., Brass, G.W., Saltzman, E., Sloan II, J., Southam, J., Whitman, J.M., 1981. Sea level variations, global sedimentation rates and the hypsographic curve. Earth Planet. Sci. Lett. 54 (1), 1–16. http://dx.doi.org/10.1016/0012-821X(81)90064-9.

Hart, M.B., 1980. A water depth model for the evolution of the planktonic foraminiferida. Nature 286, 252–254. http://dx.doi.org/10.1038/286252a0.

Hart, M.B., Smart, C.W., FitzPatrick, M.E.J., 2016. The Cretaceous/Paleogene boundary: foraminifera, sea grasses, sea level change and sequence stratigraphy. Palaeogeogr. Palaeoclimatol. Palaeoecol. 441, 420–429 (in this volume).

Hasegawa, H., Tada, R., Jiang, X., Suganuma, Y., Imsamut, S., Charusiri, P., Ichinnorov, N., Khand, Y., 2012. Drastic shrinking of the Hadley circulation during the mid-Cretaceous Supergreenhouse. Clim. Past 8, 1323–1337. http://dx.doi.org/10.5194/cp-8-1323-2012.

Hay, W.W., 1996. Tectonics and climate. Geol. Rundsch. 85 (3), 409–437. http://dx.doi.org/10.1007/BF02369000.

Hay, W.W., 2008. Evolving ideas about the Cretaceous climate and ocean circulation. Cretac. Res. 29, 725–753.

Hay, W.W., 2011. Can humans force a return to a "Cretaceous" climate? Sediment. Geol. 235, 5–26. http://dx.doi.org/10.1016/j.sedgeo.2010.04.015.

Hay, W.W., 2013. Experimenting on a Small Planet. Springer-Verlag, Berlin-Heidelberg (963 pp.).
Hay, W.W., Floegel, S., 2012. New thoughts about the Cretaceous climate and oceans. Earth Sci. Rev. 115 (4), 262–272. http://dx.doi.org/10.1016/j.earscirev.2012.09.008.
Hay, W.W., Leslie, M.A., 1990. Could possible changes in global groundwater reservoir cause eustatic sea level fluctuations? In: Geophysics Study Committee, C.o.P.S., Mathematics and Resources, National research Council (Eds.), Sea Level Change: Studies in Geophysics. National Academy Press, Washington D. C., pp. 161–170
Hay, W.W., DeConto, R.M., Wold, Ch.N., 1997. Climate: is the past the key to the future? Geol. Rundsch. 86 (2), 417–491. http://dx.doi.org/10.1007/s005310050155.
Hay, C.C., Morrow, E., Kopp, R.E., Mitrovica, J.C., 2015. Probabilistic reanalysis of twentieth-century sea-level rise. Nature 517, 481–484. http://dx.doi.org/10.1038/nature14093.
Hays, J.D., Imbrie, J., Shackleton, N.J., 1974. Variations in the Earth's orbit: Pacemaker of the ice ages. Science 194 (4270), 1121–1132. http://dx.doi.org/10.1126/science.194.4270.1121.
Hennebert, M., Robaszynski, F., Goolaerts, S., 2009. Cyclostratigraphy and chronometric scale in the Campanian–Lower Maastrichtian: the Abiod Formation at Ellès, central Tunisia. Cretac. Res. 30, 325–338.
Hilgen, F.J., Hinnov, L.A., Abdul Aziz, H., Abels, H.A., Batenburg, S., Bosmans, J.H.C., De Boer, B., Hüsing, S.K., Kuiper, K.F., Lourens, L.J., Rivera, T., Tuenter, E., Van De Wal, R.S.W., Wotzlaw, J.-F., Zeeden, C., 2015. Stratigraphic continuity and fragmentary sedimentation: the success of cyclostratigraphy as part of integrated stratigraphy. In: Smith, D.G., Bailey, R.J., Burgess, P.M., Fraser, A.J. (Eds.), Strata and Time: Probing the Gaps in Our Understanding, 10.1144/SP404.12. Geological Society, London, Special Publications 404, pp. 157–197.
Hinnov, L.A., 2013. Cyclostratigraphy and its revolutionizing applications in the earth and planetary sciences. GSA Bull. 125 (11/12), 1703–1734. http://dx.doi.org/10.1130/B30934.1.
Hinnov, L.A., Hilgen, F.J., 2012. Chapter 4: Cyclostratigraphy and Astrochronology. In: Gradstein, F.M., Ogg, J.G., Schmitz, M., Ogg, G. (Eds.), The Geologic Timescale 2012, 10.1016/B978-0-444-59425-9.00004-4 Elsevier, Amsterdam, pp. 63–83.
Holden, N.E., Bonardi, M.L., De Bièvre, P., Renne, P.R., Villa, I.M., 2011. IUPAC-IUGS common definition and convention on the use of the year as a derived unit of time. Episodes 34 (1), 39–40.
Husson, D., Galbrun, B., Laskar, J., Hinnov, L.A., Thibault, N., Gardin, S., Locklair, R.E., 2011. Astronomical calibration of the Maastrichtian (Late Cretaceous). Earth Planet. Sci. Lett. 305 (3–4), 328–340. http://dx.doi.org/10.1016/j.epsl.2011.03.008.
Immenhauser, A., 2005. High-rate sea-level change during the Mesozoic: new approaches to an old problem. Sediment. Geol. 175, 277–296.
Jacob, T., Wahr, J., Pfeffer, W.T., Swenson, S., 2012. Recent contributions of glaciers and ice caps to sea level rise. Nature 482, 514–518. http://dx.doi.org/10.1038/nature10847.
Jacobs, D.K., Sahagian, D.L., 1993. Climate induced fluctuations in sea level during non-glacial times. Nature 361, 710–712. http://dx.doi.org/10.1038/361710a0.
Jacobs, D.K., Sahagian, D.L., 1995. Milankovitch fluctuations in sea level and recent trends in sea-level change: ice may not always be the answer. In: Haq, B.U. (Ed.), Sequence Stratigraphy and Depositional Response to Eustatic, Tectonic and Climatic Forcing. Springer, Heidelberg, pp. 329–366 http://dx.doi.org/10.1007/978-94-015-8583-5.
Jarvis, I., Murphy, A.M., Gale, A.S., 2001. Geochemistry of pelagic and hemipelagic carbonates: criteria for identifying systems tracts and sea-level change. J. Geol. Soc. Lond. 158, 685–696.
Jenkyns, H.C., 2003. Evidence for rapid climate change in the Mesozoic–Palaeogene greenhouse world. Philos. Trans. R. Soc. London, Ser. A 361 (1810), 1885–1916.
Jensen, L., Rietbroek, R., Kusche, J., 2013. Land water contribution to sea level from GRACE and Jason-1 measurements. J. Geophys. Res. 118 (1), 212–226. http://dx.doi.org/10.1002/jgrc.20058.
Kaminski, M.A., Gradstein, F.M., 2005. Atlas of paleogene cosmopolitan deep-water agglutinating foraminifera. Grzybowski Found. Spec. Publ. 10, 1–574.
Kasting, J.F., Holm, N.G., 1992. What determines the volume of the oceans? Earth Planet. Sci. Lett. 109 (3–4), 507–515. http://dx.doi.org/10.1016/0012-821X(92)90110-H.
Kidder, D.L., Worsley, T.R., 2010. Phanerozoic Large Igneous Provinces (LIPs), HEATT (Haline Euxinic Acidic Thermal Transgression) episodes, and mass extinctions. Palaeogeogr. Palaeoclimatol. Palaeoecol. 295, 162–191. http://dx.doi.org/10.1016/j.palaeo.2010.05.036.
Kidder, D.L., Worsley, T.R., 2012. A human-induced hothouse climate? GSA Today 22 (2), 4–11. http://dx.doi.org/10.1130/G131A.1.
Kirschner, J.P., Kominz, M.A., Mwakanyamale, K.E., 2010. Quantifying extension of passive margins: implications for sea level change. Tectonics 29 (4), TC4006. http://dx.doi.org/10.1029/2009TC002557.
Kominz, M.A., 1984. Ocean ridge volumes and sea-level change—an error analysis. In: Schlee, J.S. (Ed.), Interregional Unconformities and Hydrocarbon Accumulation. American Association of Petroleum Geologists Memoir 36, pp. 109–127.
Kominz, M.A., Browning, J.V., Miller, K.G., Sugarman, P.J., Misintseva, S., Scotese, C.R., 2008. Late Cretaceous to Miocene sea-level estimates from the New Jersey and Delaware coastal plain coreholes: an error analysis. Basin Res. 20, 211–226.
Konikow, L.F., 2011. Contribution of global groundwater depletion since 1900 to sea-level rise. Geophys. Res. Lett. 38 (17), L17401. http://dx.doi.org/10.1029/2011gl048604.
Korenaga, J., 2011. Thermal evolution with a hydrating mantle and the initiation of plate tectonics in the early Earth. J. Geophys. Res. 116, B12403. http://dx.doi.org/10.1029/2011JB008410.
Koutsoukos, E.A.M., Hart, M.B., 1990. Cretaceous foraminiferal morphogroup distribution patterns, palaeocommunities and trophic structures: a case study from the Sergipe Basin, Brazil. Trans. R. Soc. Edinb. Earth Sci. 81, 221–246.
Kuiper, K.F., Deino, A., Hilgen, F.J., Krijgsman, W., Renne, P.R., Wijbrans, J.R., 2008. Synchronizing rock clocks of Earth history. Science 320, 500–504.
Laskar, J., Fienga, A., Gastineau, M., Manche, H., 2011. La2010: a new orbital solution for the long-term motion of the Earth. Astron. Astrophys. 532, A89. http://dx.doi.org/10.1051/0004-6361/201116836.
Li, L., Keller, G., Adatte, T., Stinnesbeck, W., 2000. Late Cretaceous sea level changes in Tunisia: a multi-disciplinary approach. J. Geol. Soc. Lond. 157, 447–458.
Liu, Y., Peltier, W.R., 2013. Sea level variations during snowball Earth formation: 1. A preliminary analysis. J. Geophys. Res. Solid Earth 118, 4410–4424. http://dx.doi.org/10.1002/jgrb.50293.
Locklair, R.E., Sageman, B.B., 2008. Cyclostratigraphy of the Upper Cretaceous Niobrara Formation, Western Interior, U.S.A.: a Coniacian–Santonian orbital timescale. Earth Planet. Sci. Lett. 269, 540–553.
Lovell, B., 2010. A pulse in the planet: regional control of high-frequency changes in relative sea level by mantle convection. J. Geol. Soc. Lond. 167, 637–648. http://dx.doi.org/10.1144/0016-76492009-127.
MacLeod, K.G., Huber, B.T., Jiménez Berrocoso, Á., Wendler, I., 2013. A stable and hot Turonian without glacial $\delta^{18}O$ excursions is indicated by exquisitely preserved Tanzanian foraminifera. Geology 41, 1083–1086.
Marine rapid environmental/climatic change in the cretaceous greenhouse world. In: Hu, X., Wagreich, M., Yilmaz, I.O. (Eds.), Cretac. Res. 38, 1–112.
Martinez, M., Deconinck, J.-F., Pellenard, P., Reboulet, S., Riquier, L., 2013. Astrochronology of the Valanginian Stage from reference sections (Vocontian Basin, France) and palaeoenvironmental implications for the Weissert Event. Palaeogeogr. Palaeoclimatol. Palaeoecol. 376, 91–102. http://dx.doi.org/10.1016/j.palaeo.2013.02.021.
Martinez, M., Deconinck, J.-F., Pellenard, P., Riquier, L., Company, M., Reboulet, S., Moiroud, M., 2015. Astrochronology of the Valanginian–Hauterivian stages (Early Cretaceous): chronological relationships between the Paraná–Etendeka large igneous province and the Weissert and the Faraoni events. Glob. Planet. Chang. 131, 158–173. http://dx.doi.org/10.1016/j.gloplacha.2015.06.001.
Miller, K.G., Kominz, M.A., Browning, J.V., Wright, J.D., Mountain, G.S., Katz, M.E., Sugarman, P.J., Cramer, B.S., Christie-Blick, N., Pekar, S.F., 2005a. The Phanerozoic record of global sea-level change. Science 310, 1293–1298.
Miller, K.G., Wright, J.D., Browning, J.V., 2005b. Visions of ice sheets in a greenhouse world. Mar. Geol. 217, 215–231.
Miller, K.G., Mountain, G.S., Wright, J.D., Browning, J.V., 2011. A 180-million-year record of sea level and ice volume variations from continental margin and deep-sea isotopic records. Oceanography 24 (2), 40–53. http://dx.doi.org/10.5670/oceanog.2011.26.
Miller, K.G., Wright, J.D., Katz, M.E., Wade, B.S., Browning, J.V., Cramer, B.S., Rosenthal, Y., 2009. Climate threshold at the Eocene-Oligocene transition: Antarctic ice sheet influence on ocean circulation. In: Koeberl, C., Montanari, A. (Eds.), The Late Eocene Earth—Hothouse, Icehouse, and Impacts. Geol. Soc. Am. Spec. Pap 452, pp. 169–178. http://dx.doi.org/10.1130/2009.2452(11).
Milly, P.C.D., Cazenave, A., Famiglietti, J.S., Gornitz, V., Laval, K., Lettenmaier, D.P., Sahagian, D.L., Wahr, J.M., Wilson, C.R., 2010. Terrestrial water-storage contributions to sea-level rise and variability. In: Church, J.A., Woodworth, P.L., Aarup, T., Wilson, W.S. (Eds.), Understanding Sea-Level Rise and Variability, 1st ed. Wiley-Blackwell, Chichester, pp. 226–255.
Milne, G.A., Mitrovica, J.X., 1998. Postglacial sea-level change on a rotating Earth. Geophys. J. Int. 133 (1), 1–19. http://dx.doi.org/10.1046/j.1365-246X.1998.1331455.x.
Milne, G.A., Mitrovica, J.X., 2008. Searching for eustasy in deglacial sea-level histories. Quat. Sci. Rev. 27 (25–26), 2292–2302. http://dx.doi.org/10.1016/j.quascirev.2008.08.018.
Mitchum, G.T., Nerem, S.R., Merrifield, M.A., Gehrels, W.R., 2010. Chapter 5 — Modern sea-level-change estimates. In: Church, J.A., Woodworth, P.L., Aarup, T., Wilson, W.S. (Eds.), Understanding Sea-Level Rise and Variability, 1st ed. Wiley-Blackwell, Chichester, pp. 122–142.
Mitrovica, J.X., Milne, G.A., 2003. On post-glacial sea level: I. General theory. Geophys. J. Int. 154, 253–267. http://dx.doi.org/10.1046/j.1365-246X.2003.01942.x.
Mitrovica, J.X., Peltier, W.R., 1991. On postglacial geoid subsidence over the equatorial oceans. J. Geophys. Res. Solid Earth 96 (B12), 20053–20071. http://dx.doi.org/10.1029/91JB01284.
Mitrovica, J.X., Tamisiea, M.E., Davis, J.L., Milne, G.A., 2001. Recent mass balance of polar ice sheets inferred from patterns of global sea-level change. Nature 409 (6823), 1026–1029. http://dx.doi.org/10.1038/35059054.
Moore, C.H., Wade, W.J., 2013. Carbonate Reservoirs — Porosity and Diagenesis in a Sequence Stratigraphic Framework. Dev. Sedimentol 67, 1–374. http://dx.doi.org/10.1016/B978-0-444-53831-4.09992-4.
Moriya, K., Wilson, P.A., Friedrich, O., Erbacher, J., Kawahata, H., 2007. Testing for ice sheets during the mid-Cretaceous greenhouse using glassy foraminiferal calcite from the mid-Cenomanian tropics on Demerara Rise. Geology 35, 615–618.
Moucha, R., Forte, A.M., Mitrovica, J.X., Rowley, D.B., Quere, S., Simmons, N.A., Grand, S.P., 2008. Dynamic topography and long-term sea-level variations: there is no such thing as a stable continental platform. Earth Planet. Sci. Lett. 271 (1–4), 101–108. http://dx.doi.org/10.1016/j.epsl.2008.03.056.
Müller, R.D., Sdrolias, M., Gaina, C., Steinberger, B., Heine, C., 2008. Long-term sea-level fluctuations driven by ocean basin dynamics. Science 319, 1357–1362. http://dx.doi.org/10.1126/science.1151540.
Murray, J., 1991. Ecology and Paleoecology of Benthic Foraminifera. Longman Scientific & Technical, London (398 pp.).
Nicholls, R.J., 2010. Impacts of and responses to sea-level rise. In: Church, J.A., Woodworth, P.L., Aarup, T., Wilson, W.S. (Eds.), Understanding Sea-Level Rise and Variability, 1st ed. Wiley-Blackwell, Chichester, pp. 17–51.
Nicholls, R.J., Cazenave, A., 2010. Sea-level rise and its impact on coastal zones. Science 328, 1517–1520. http://dx.doi.org/10.1016/10.1126/science.1185782.
Nielsen, K., Khan, S.A., Korsgaard, N.J., Kjær, K.H., Wahr, J., Bevis, M., Stearns, L.A., Timm, L.H., 2012. Crustal uplift due to ice mass variability on Upernavik Isstrøm, west

Greenland. Earth Planet. Sci. Lett. 353–354, 182–189. http://dx.doi.org/10.1016/j.epsl.2012.08.024.
Ogg, G.M., 2012. Geomagnetic Polarity Time Scale. In: Gradstein, F.M., Ogg, J.G., Schmitz, M.D., Ogg, G.M (Eds.), The Geologic Time Scale 2012. Elsevier, Amsterdam, pp. 85–113 http://dx.doi.org/10.1016/B978-0-444-59425-9.00005-6.
Olde, K., Jarvis, I., Uličný, D., Pearce, M.A., Trabucho-Alexandre, J., Čech, S., Gröcke, D.R., Laurin, J., Švábenická, L., Tocher, B.A., 2015. Geochemical and palynological sea-level proxies in hemipelagic sediments: a critical assessment from the Upper Cretaceous of the Czech Republic. Palaeogeogr. Palaeoclimatol. Palaeoecol. 435, 222–243. http://dx.doi.org/10.1016/j.palaeo.2015.06.018.
Petersen, K.D., Nielsen, S.B., Clausen, O.R., Stephenson, R., Gerya, T., 2010. Small-scale mantle convection produces stratigraphic sequences in sedimentary basins. Science 329, 827–830. http://dx.doi.org/10.1126/science.1190115.
Piecuch, C.G., Ponte, R.M., 2014. Mechanisms of global-mean steric sea level change. J. Clim. 27, 824–834. http://dx.doi.org/10.1175/JCLI-D-13-00373.1.
Pitman, W.C., 1978. Relationship between eustasy and stratigraphic sequences of passive margins. Geol. Soc. Am. Bull. 89, 1389–1403.
Pokhrel, Y.N., Hanasaki, N., Yeh, P.J.F., Yamada, T.J., Kanae, S., Oki, T., 2012. Model estimates of sea-level change due to anthropogenic impacts on terrestrial water storage. Nat. Geosci. 5 (6), 389–392. http://dx.doi.org/10.1038/ngeo1476.
Price, G.D., Nunn, E.V., 2010. Valanginian isotope variation in glendonites and belemnites from Arctic Svalbard: transient glacial temperatures during the Cretaceous greenhouse. Geology 38, 251–254.
Rahmstorf, S., Box, J.E., Feulner, G., Mann, M.E., Robinson, A., Rutherford, S., Schaffernicht, E.J., 2015. Exceptional twentieth-century slowdown in Atlantic Ocean overturning circulation. Nat. Clim. Chang. 5, 475–480. http://dx.doi.org/10.1038/nclimate2554.
Robson, J., Hodson, D., Hawkins, E., Sutton, R., 2014. Atlantic overturning in decline? Nat. Geosci. 7, 2. http://dx.doi.org/10.1038/ngeo2050.
Rose, B.E.J., Ferreira, D., 2013. Ocean heat transport and water vapor greenhouse in a warm equable climate: a new look at the low gradient paradox. J. Clim. 26, 2117–2136. http://dx.doi.org/10.1175/JCLI-D-11-00547.1.
Ruban, D.A., Zorina, S.O., Conrad, C.P., 2010. No global-scale transgressive-regressive cycles in the Thanetian (Paleocene): evidence from interregional correlation. Palaeogeogr. Palaeoclimatol. Palaeoecol. 295, 226–235.
Sageman, B.B., Meyers, S.R., Arthur, M.A., 2006. Orbital time scale and new C-isotope record for Cenomanian-Turonian boundary stratotype. Geology 34, 125–128. http://dx.doi.org/10.1130/G22074.1.
Sageman, B.B., Singer, B.S., Meyers, S.R., Siewert, S.E., Walaszczyk, I., Condon, D.J., Jicha, B.R., Obradovich, J.D., Sawyer, D.A., 2014. Integrating 40Ar/39Ar, U-Pb, and astronomical clocks in the Cretaceous Niobrara Formation, Western Interior Basin, USA. Geol. Soc. Am. Bull 126 (7–8), 956–973. http://dx.doi.org/10.1130/B30929.1.
Saltzman, M.R., Thomas, E., 2012. Carbon isotope stratigraphy. In: Gradstein, F.M., Ogg, J.G., Schmitz, M.D., Ogg, G.M (Eds.), The Geologic Time Scale 2012. Elsevier, Amsterdam, pp. 207–232 http://dx.doi.org/10.1016/B978-0-444-59425-9.00011-1.
Sames, B., 2015. Integrating palaeoenvironmental and climate cyclicities — optimizing the stratigraphic framework in the non-marine Lower Cretaceous. Geophys. Res. Abstr. 17 (http://meetingorganizer.copernicus.org/EGU2015/EGU2015-1908.pdf).
Sames, B., Horne, D.J., 2012. Latest Jurassic to Cretaceous non-marine ostracod biostratigraphy: unde venis, quo vadis? J. Stratigr. 36 (2), 267–290.
Sandu, C., Lenardic, A., McGovern, P., 2011. The effects of deep water cycling on planetary thermal evolution. J. Geophys. Res. 116, B12404. http://dx.doi.org/10.1029/2011JB008405.
Şengör, A.M.C., 2015. The dounder of modern geology died 100 years ago: the scientific work and legacy of Eduard Suess. Can. Geosci. 42, 181–246. http://dx.doi.org/10.12789/geocanj.2015.42.070.
Shackleton, N.J., Kennett, J.P., 1975. Paleotemperature history of the Cenozoic and initiation of Antarctic glaciation: oxygen and carbon isotopic analyses in DSDP Sites 277, 279, and 281. Initial Rep. Deep Sea Drill. Proj. 29, 743–755.
Shennan, I., 2015. Chapter 2 — Handbook of sea-level research: framing research questions. In: Shennan, I., Long, A.J., Horton, B.P. (Eds.), Handbook of Sea-Level Research, 1st ed. John Wiley and Sons, Chichester, pp. 2–25.
Shennan, I., Horton, B., 2002. Holocene land and sea-level changes in Great Britain. J. Quat. Sci. 17 (5–6), 511–526 (http://dx.doi.org/10.1002/jqs.710).
Shennan, I., Milne, G., Bradley, S., 2012. Late Holocene vertical land motion and relative sea-level changes: lessons from the British Isles. J. Quat. Sci. 27 (1), 64–70. http://dx.doi.org/10.1002/jqs.1532.
Simmons, M., 2011. The ups and downs of eustatic sea level change. GEO ExPro 8 (5), 64–69 (http://www.geoexpro.com/magazine/vol-8-no-5).
Simmons, M.D., 2012. Sequence stratigraphy and sea-level change. In: Gradstein, F.M., Ogg, J.G., Schmitz, M.D., Ogg, G.M. (Eds.), The Geologic Time Scale 2012. Elsevier, Amsterdam, pp. 239–267.
Skelton, P.W., 2003. The Cretaceous World. Cambridge Univ Press, London (360 pp.).
Slangen, A.B.A., van de Wal, R.S.W., Wada, Y., Vermeersen, L.L.A., 2014. Comparing tide gauge observations to regional patterns of sea-level change (1961-2003). Earth Syst. Dyn. 5 (1), 243–255 (http://dx.doi.org/10.5194/esd-5-243-2014).
Sliter, W.V., Baker, R.A., 1972. Cretaceous bathymetric distribution of benthic foraminifers. J. Foraminifer. Res. 2 (4), 167–183 (http://dx.doi.org/10.2113/gsjfr.2.4.167).
Spasojevic, S., Gurnis, M., 2012. Sea level and vertical motion of continents from dynamic Earth models since the Late Cretaceous. Am. Assoc. Pet. Geol. Bull. 96 (11), 2037–2064. http://dx.doi.org/10.1306/03261211121.
Sprovieri, M., Sabatino, N., Pelosi, N., Batenburg, S.J., Coccioni, R., Iavarone, M., Mazzola, S., 2013. Late Cretaceous orbitally-paced carbon isotope stratigraphy from the Bottaccione Gorge (Italy). Palaeogeogr. Palaeoclimatol. Palaeoecol. 379–380, 81–94.
Steffen, K., Thomas, R.H., Rignot, E., Cogley, J.G., Dyurgerov, M.B., Raper, S.C.B., Huybrechts, P., Hanna, E., 2010. Cryospheric contributions to sea-level rise and variability. In: Church, J.A., Woodworth, P.L., Aarup, T., Wilson, W.S. (Eds.), Understanding Sea-Level Rise and Variability, 1st ed. Wiley-Blackwell, Chichester, pp. 177–225.
Stoll, H.M., Schrag, D.P., 1996. Evidence for glacial control of rapid sealevel changes in the Early Cretaceous. Science 272, 1771–1774.
Suarez, M.B., González, L.A., Ludvigson, G.A., 2011. Quantification of a greenhouse hydrologic cycle from equatorial to polar latitudes: the mid-Cretaceous water bearer revisited. Palaeogeogr. Palaeoclimatol. Palaeoecol. 307, 301–312. http://dx.doi.org/10.1016/j.palaeo.2011.05.027.
Suess, E., 1888. Das Antlitz der Erde. part 3: Die Meere der Erde. F. Tempsky, Wien. Vol. 2 (703 pp.).
Syvitski, J.P.M., Kettner, A.J., Overeem, I., Hutton, E.W.H., Hannon, M.T., Brakenridge, G.R., Day, J., Vörösmarty, C., Saito, Y., Giosan, L., Nicholls, R.J., 2009. Sinking deltas due to human activities. Nat. Geosci. 2, 681–686. http://dx.doi.org/10.1038/ngeo629.
Velinga, M., Wood, R.A., 2002. Global climatic impacts of a collapse of the Atlantic thermohaline circulation. Clim. Chang. 54, 251–267. http://dx.doi.org/10.1023/A:1016168827653.
Voigt, S., Schönfeld, J., 2010. Cyclostratigraphy of the reference section for the Cretaceous white chalk of northern Germany, Lägerdorf–Kronsmoor: a late Campanian–early Maastrichtian orbital time scale. Palaeogeogr. Palaeoclimatol. Palaeoecol 287, 67–80. http://dx.doi.org/10.1016/j.palaeo.2010.01.017.
Wada, Y., van Beek, L.P.H., van Kempen, C.M., Reckman, J.W.T.M., Vasak, S., Bierkens, M.F.P., 2010. Global depletion of groundwater resources. Geophys. Res. Lett. 37 (20), L20402. http://dx.doi.org/10.1029/2010GL044571.
Wada, Y., van Beek, L.P.H., Sperna Weiland, F.C., Chao, B.F., Wu, Y.-H., Bierkens, M.F.P., 2012. Past and future contribution of global groundwater depletion to sea-level rise. Geophys. Res. Lett. 39 (9), L09402. http://dx.doi.org/10.1029/2012gl051230.
Wagreich, M., 2009. Stratigraphic constraints on climate control of Lower Cretaceous oceanic red beds in the Northern Calcareous Alps (Austria). In: Hu, X., Wang, C., Scott, R.W., Wagreich, M., Jansa, L. (Eds.), Cretaceous Oceanic Red Beds: Stratigraphy, Composition, Origins, and Paleoceanographic and Paleoclimatic Significance. SEPM Special Publication 91, pp. 91–98.
Wagreich, M., Hohenegger, J., Neuhuber, S., 2012. Nannofossil biostratigraphy, strontium and carbon isotope stratigraphy, cyclostratigraphy and an astronomically calibrated duration of the Late Campanian *Radotruncana calcarata* Zone. Cretac. Res. 38, 80–96.
Wagreich, M., Lein, R., Sames, B., 2014. Eustasy, its controlling factors, and the limnoeustatic hypothesis — concepts inspired by Eduard Suess. J. Aust. Earth Sci. 107 (1), 115–131 (http://www.univie.ac.at/ajes/archive/volume_107_1/wagreich_et_al_ajes_107_1.pdf).
Waltham, D., 2015. Milankovitch period uncertainties and their impact on cyclostratigraphy. J. Sediment. Res. 85 (8), 990–998.
Wang, C., Huang, Y., Zhao, X., 2009. Unlocking a Cretaceous geologic and geophysical puzzle: scientific drilling of Songliao Basin in northeast China. Lead. Edge 28 (3), 340–344.
Wendler, J.E., Wendler, I., 2016. What drove cyclic sea-level fluctuations during the mid-Cretaceous greenhouse climate? Palaeogeogr. Palaeoclimatol. Palaeoecol. 441, 412–419 (in this volume).
Wendler, J.E., Meyers, S.R., Wendler, I., Vogt, C., Kuss, J., 2011. Drivers of cyclic sea level changes during the Cretaceous greenhouse: a new perspective from the Levant Platform. Geol. Soc. Am. Abstr. Programs 43 (5), 376.
Wendler, J.E., Meyers, S.R., Wendler, I., Kuss, J., 2014. A million-year-scale astronomical control on Late Cretaceous sea-level. Newsl. Stratigr. 47 (1), 1–19. http://dx.doi.org/10.1127/0078-0421/2014/0038.
Wendler, J.E., Wendler, I., Vogt, C., Kuss, J., 2016. Link between cyclic eustatic sea-level change and continental weathering: evidence for aquifer-eustasy in the Cretaceous. Palaeogeogr. Palaeoclimatol. Palaeoecol. 441, 430–437 (in this volume).
Wetzel, R.G., 2001. Limnology: Lake and River Ecosystems. 3rd ed. Academic Press, London (1006 pp.).
Widmark, J.G.V., Speijer, R.P., 1997. Benthic foraminiferal faunas and trophic regimes at the terminal Cretaceous Tethyan seafloor. Palaios 12 (4), 354–371. http://dx.doi.org/10.2307/3515335.
Wilmsen, M., Nagm, E., 2013. Sequence stratigraphy of the lower Upper Cretaceous (Upper Cenomanian–Turonian) of the Eastern Desert, Egypt. Newsl. Stratigr. 46 (1), 23–46.
Wissler, L., Weissert, H., Buonocunto, F.P., Ferreri, V., D'Argenio, B., 2004. Calibration of the Early Cretaceous time scale: a combined chemostratigraphic and cyclostratigraphic approach to the Barremian-Aptian interval Campania Apennines and southern Alps (Italy). In: D'Argenio, B., Fischer, A.G., Premoli Silva, I., Weissert, H., Ferreri, V. (Eds.), Cyclostratigraphy, Approaches and Case Histories. SEPM Special Publication 81, pp. 123–134.
Wolfgring, E., Hohenegger, J., Wagreich, M., 2016. Assessing pelagic palaeoenvironments using foraminiferal assemblages — a case study from the late Campanian *Radotruncana calcarata* Zone (Upper Cretaceous, Austrian Alps). Palaeogeogr. Palaeoclimatol. Palaeoecol. 441, 467–492 (in this volume).
Woodworth, P.L., Menéndez, M., 2015. Changes in the mesoscale variability and in extreme sea levels over two decades as observed by satellite altimetry. J. Geophys. Res. Oceans 120, 64–77. http://dx.doi.org/10.1002/2014JC010363.
Woodworth, P.L., White, N.J., Jevrejeva, S., Holgate, S.J., Church, J.A., Gehrels, W.R., 2009. Evidence for the accelerations of sea level on multi-decade and century timescales. Int. J. Climatol. 29, 777–789. http://dx.doi.org/10.1002/joc.1771.
Wu, H., Zhang, S., Jiang, G., Yang, T., Guo, J., Li, H., 2013. Astrochronology for the Early Cretaceous Jehol Biota in northeastern China. Palaeogeogr. Palaeoclimatol. Palaeoecol. 385, 221–228. http://dx.doi.org/10.1016/j.palaeo.2013.05.017.
Xu, X.Q., Lithgow-Bertelloni, C., Conrad, C.P., 2006. Global reconstructions of Cenozoic sea floor ages: implications for bathymetry and sea level. Earth Planet. Sci. Lett. 243 (3–4), 552–564. http://dx.doi.org/10.1016/j.epsl.2006.01.010.

Yilmaz, I.O., Altiner, D., 2001. Use of sedimentary structures in the recognition of sequence boundaries in the Upper Jurassic (Kimmeridgian) – Upper Cretaceous (Cenomanian) peritidal carbonates of the Fele (Yassıbel) area (Western Taurides, Turkey). Int. Geol. Rev. 43 (8), 736–754.
Yilmaz, I.O., Altiner, D., 2006. Cyclic palaeokarst surfaces in Aptian peritidal carbonate successions (Taurides, southwest Turkey): internal structure and response to mid-Aptian sea-level fall. Cretac. Res. 27, 814–827.
Zorina, S.O., 2014. Sedimentation regime and accommodation space in the Middle Jurassic–Lower Cretaceous on the eastern Russian Plate. Russ. Geol. Geophys. 55, 1195–1204. http://dx.doi.org/10.1016/j.rgg.2014.09.005.
Zorina, S.O., Dzyuba, O.S., Shurygin, B.N., Ruban, D.A., 2008. How global are the Jurassic–Cretaceous unconformities? Terra Nova 20, 341–346.