**Hiatuses and condensation: an estimation of time lost on a shallow carbonate platform**

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**ABSTRACT**

On shallow carbonate platforms, the sedimentary record is highly fragmentary because low accommodation commonly leads to non-deposition, erosion, reworking and condensation. Consequently, it is difficult to quantify the time that is actually recorded and to estimate sedimentation rates. The Berriasian platform in the Swiss and French Jura Mountains offers an opportunity to address this problem. Facies evolution through time displays deepening-shallowing trends of different orders, resulting in a hierarchical stacking of sequences. The sequence- and cyclostratigraphic analysis relates these sequences to orbital cycles with durations of 20, 100, and 400 kyr. Many surfaces in the record of shallow-water carbonates can be interpreted as hiatuses due to non-deposition in the supratidal realm, reactivation of shoals or scouring by currents. Strongly bioturbated intervals and hardgrounds commonly result from sediment starvation. These same processes and products can be observed on modern carbonate platforms, and the duration of the hiatuses or condensed intervals can be inferred (a few hours to a few hundreds of years). Amalgamations and narrow stacking of such surfaces create sequence-boundary zones that can be followed over the entire platform. Time lost or condensed in these zones may correspond to hundreds of thousands of years. Time distribution in the studied sections thus is highly irregular: short periods of sedimentation alternated with long periods of non-deposition, erosion, reworking, and condensation. Furthermore, due to substrate morphology and laterally variable depositional environments, the depositional record of a given time interval varied significantly across the platform. In one example, it can be estimated that – applying modern sedimentation rates – a sedimentary record spanning 800 kyr could have been deposited within 44 kyr if there were no hiatuses or condensations. It becomes clear that, when estimating carbonate production and sediment accumulation rates in ancient records, the shortest possible time spans must be used to minimize the effect of time lost in hiatal surfaces and condensed intervals.

**INTRODUCTION**

Faithful reconstruction of Earth’s history depends on the completeness of the rock record. Chronostratigraphic charts are established by radiometrically and/or astrochronologically dating rocks that represent this record. No place on Earth contains the entire record from the Hadean to the Holocene, and completeness is attempted by patching fragmentary records together at a global scale. One such example is the IUGS International Chronostratigraphic Chart that is constantly updated following new findings and improved dating methods (www.stratigraphy.org).

Relatively complete records covering parts of Earth history are obtained from sediments that have been deposited at high rates in regularly subsiding oceanic or lacustrine basins below wave base and out of the influence of bottom currents, and where bioturbation is minimal due to low oxygen levels. Thus, time windows are opened in which the sedimentary, hydrological, climatic
and biological evolution in a specific basin can be reconstructed. An example for a well-preserved marine record is the Miocene to Holocene sediment stack of the Mediterranean basin, which reflects orbital cyclicity (Fischer et al., 2009) and where specific intervals can be interpreted with a millennial- to centennial-scale resolution (Rodrigo-Gámiz et al., 2014). For lake sediments, a good example is the 600 kyr Pleistocene to Holocene record drilled and interpreted in Lake Van, Turkey (Stockhecke et al., 2014).

When it comes to shallow sedimentary environments, gaps in the depositional record are the norm. If such an environment lies on land above sea or lake level, no sediment at all may be deposited and the corresponding time interval leaves no trace, or else erosion of the terrestrial sediment may partly or completely destroy the record. In the subaqueous zone, the sediment may be scour and reworked by currents or overturned by biogenic activity, and the final record does not anymore reflect the original depositional history. This incompleteness of the depositional record has been recognized and discussed by many authors. For example, Barrell (1917) wrote that most stratigraphic records are characterized by diastems at bedding planes. In the case of tide-influenced shallow siliciclastic sediments, Reineck (1960) estimated that only 10^4 to 10^5 of geologic time is actually preserved in the sedimentary column. Ager (1980) coined the phrase that the ‘stratigraphical record is a lot of holes tied together with sediment’. Sadler (1981) wrote that ‘sedimentation is an essentially discontinuous process’ and compared sediment accumulation rates with the time span over which this accumulation took place. Dott (1983) discussed the episodic nature of sedimentation, and Miall (2014) wrote about the ‘emptiness of the stratigraphic record’. Different types of condensation processes on shallow water platforms were illustrated by Gómez & Fernández-López (1994). Kemp (2012) modelled the distribution of hiatuses in the stratigraphic record, and Hill et al. (2012) focused on the preservation potential of shallow-water carbonate sediments exposed to high-frequency, orbitally controlled sea-level fluctuations. Good examples of complex discontinuity surfaces have been published by Sattler et al. (2005) and Rameil et al. (2012) from the Cretaceous carbonate platforms in Oman, Waite et al. (2013) from the Late Jurassic in the Swiss Jura Mountains, and Brlek et al. (2014) from the Cretaceous-Palaeogene boundary in Croatia.

Figure 1 schematically illustrates how different processes on a shallow carbonate platform lead to the sedimentary record. After some marine, subtidal sediment accumulation, a drop in relative sea-level leads to erosion that removes part of the originally deposited sediment. The sediment bed is then cemented in the freshwater vadose zone, and pedogenesis sets in. During subaerial exposure, some biological and mechanical erosion goes on but is counterbalanced by deposition, so that a thin bed of soil with limestone clasts is formed. Rapid marine transgression then picks up some of these clasts and incorporates them at the base of a bed that shows cross-bedding of an inter- to subtidal dune. Incipient hardgrounds partly consolidate the sediment, which is reworked by currents and redeposited in a marly matrix. A phase of subtidal carbonate deposition sets in, incorporating some reworked mud-clasts. The centre of the bed is characterized by intense bioturbation, implying a lowered sedimentation rate (i.e. condensation of time in a thin stratigraphic interval). The top of the bed is a hardground that is bored and encrusted by microorganisms, and mineralized. There is a balance between biological erosion and accumulation. This hypothetical history demonstrates how unequally time is distributed in the depositional record, but also that there is a potential to decipher some of this history by careful analysis.

The purpose of the present paper is to document and interpret the incompleteness of the depositional record of a shallow, subtropical carbonate platform of Berriasian (Early Cretaceous) age. It will be attempted not only to estimate the amount of time lost in non-deposition, erosion, and condensation but also to seek for the causes of the gaps and the processes that led to the final rock record as observed today in the studied outcrops. Furthermore, sedimentation rates will be discussed. The goal is to evaluate to what extent a fragmentary depositional record can be used to interpret the full history of a carbonate platform.

GEOPHYSIC, PALAEOGEOGRAPHIC AND STRATIGRAPHIC SETTING

The present study is based on the analysis of 35 sections in the Swiss and French Jura Mountains that have been logged at centimetre-scale (Strasser, 1988; Pasquier, 1995; Hillgärtner, 1999; Tresch, 2007) (Fig. 2). They represent the shallow-water carbonate realm during the Berriasian. The Jura platform was situated between the Paris Basin and the Helvetic Shelf, at a paleoaltitude of 27 to 28°N (Dercourt et al., 2000) (Fig. 3). The climate was subtropical and the potential for organic carbonate production was high. Episodically, quartz grains, clays, and nutrients were washed in from the emergent Hercynian massifs. The platform was block-faulted and structured into highs and shallow depressions, creating a complex substrate morphology (Allenbach, 2002).

The studied interval covers three formations (Fig. 4): the upper part of the Twannbach Formation (Vouglans Member; Bernier, 1984), the Goldberg Formation (Häfeli,
and the Pierre-Châtel Formation (Steinhauser & Lombard, 1969). Biostratigraphic dating is given by ammonites and benthic foraminifera (Clavel et al., 1986), and by charophytes and charophyte-ostracode assemblages (Mojon, 2002). Within the frame of this biostratigraphy, the large-scale sequence boundaries recognized in the studied outcrops can be correlated with those of the sequence-chronostratigraphic chart of Hardenbol et al. (1998).

A sequence-stratigraphic and cyclostratigraphic interpretation has been proposed by Pasquier & Strasser (1997), Strasser & Hillgärtner (1998), and Strasser et al. (2004). According to the chart of Hardenbol et al. (1998), the time span between sequence boundaries Be 1 and Be 4 is 3-2 Myr (Fig. 4). In the same interval, 32 small-scale depositional sequences are counted (for definition of such sequences see below), suggesting that one small-scale sequence corresponds to the short eccentricity cycle of 100 kyr (Berger et al., 1989). The fact that the ages of sequence boundaries Be 2 and Be 3 do not correspond to our cyclostratigraphic interpretation can be explained by the physical expression of prominent large-scale boundaries that does not necessarily happen at the same time on the platform and in the basin on which the chart of Hardenbol et al. (1998) is based (Strasser et al., 2000). Between sequence boundary Be 4 and the base of the Pierre-Châtel Formation, sedimentation on the shallow platform was much reduced but continuous in the Vocontian Basin, where the cyclostratigraphic analysis of hemipelagic limestone-marl alternations suggests that this interval lasted about 900 kyr (Strasser et al., 2004). The Pierre-Châtel Formation itself is composed of nine small-scale sequences where it is fully developed, but non-deposition and/or erosion at sequence boundary Be 5 locally cut off the topmost sequences (Pasquier, 1995; Strasser et al., 2004). The total time interval between Be 4 and Be 5 would thus have lasted about 1-8 Myr, which is close to the 1-7 Myr mentioned in the chart of Hardenbol.

Fig. 1. Hypothetical sedimentary record on a shallow, carbonate-dominated platform, displaying typical features such as palaeosol, reworked lithoclasts, bioturbation and hardground. Hypothetical sediment accumulation and erosion rates are plotted along a time axis that represents a few thousand years. Note that erosion and sedimentation may occur simultaneously during soil and hardground formation, albeit at low rates. Erosion includes mechanical, chemical and biological processes. Sediment accumulation includes in situ carbonate production, import of particles by currents or wind, and biogenic encrustation on the hardground surface. For discussion refer to text.
et al. (1998). It has to be noted that the ages of the sequence boundaries published by Hardenbol et al. (1998) derive from interpolation between the ages of the lower and upper boundaries of the Berriasian stage as given by Gradstein et al. (1995: 144.2 ± 2.6 and 137 ± 2.2 Ma, respectively). The newest chronostratigraphic chart (www.stratigraphy.org, 2015/01) indicates 145.0 ± 4.0 and 139.8 ± 3.0 Ma for these limits, thus reducing the duration of the Berriasian stage from ≈7.2 to ≈5.2 Myr (but increasing the error margins). Nevertheless, because the chart of Hardenbol et al. (1998) also compiles the biostratigraphy by which the studied formations are calibrated, the values of this older chart are retained.

METHODS

The sections (Fig. 2) have been logged at centimetre-scale and densely sampled. Thin-sections were prepared for the rock samples; marls were washed and the residue picked for microfossils. Under the optical microscope or the binocular, microfossils have been analysed using the Dunham (1962) classification and a semi-quantitative estimation of the abundance of rock constituents. Special attention has been paid to sedimentary structures and to discontinuity surfaces (Clari et al., 1995; Hillgartner, 1998). The sum of this sedimentological information was then used to interpret the depositional environments.

For the stable-isotope analyses (O18/O16 and C13/C12), 5 to 10 mg of powdered bulk rock were reacted with 100% H3PO4 at 75°C and analysed in a Finnigan Mat 252 mass spectrometer at the University of Erlangen. The ratios are reported in ‰ relative to the Vienna Pee Dee Belemnite (PDB) standard. Mean standard deviation was less than 0.1‰ for δ13C and δ18O. If possible, facies rich in micritic matrix and poor in cements were chosen in order to obtain an average signal with minimal late-diagenetic influence.

For the sequence-stratigraphic interpretation of the facies evolution, the nomenclature of Vail et al. (1991) is
sequences are typically thinner around the small-scale and medium-scale sequence boundaries, which suggests reduced accommodation space. Thick elementary sequences in the central parts of these sequences imply higher accommodation space. The medium-scale sequences then group into large-scale sequences, which can be compared with the sequences of Hardenbol et al. (1998) (Fig. 4). In some cases, no unique sequence boundaries or maximum-flooding surfaces can be identified; there, sequence-boundary zones covering intervals of lowest accommodation and maxi-

Fig. 4. Stratigraphy of the early, middle and the lower part of the late Berriasian. Sequence chronostratigraphy (ages of sequence boundaries Be 1 to Be 5 in Ma) and ammonite zones and subzones according to Hardenbol et al. (1998). The two benthic foraminifera of stratigraphic value are coeval to the Paramimounum subzone (Clavel et al., 1986). Charophyte biostratigraphy and the zones based on charophyte-ostracode assemblages are from Mojon (2002). The Goldberg and Pierre-Châtel formations are tied to the chronostratigraphy via the available fossil content (Clavel et al., 1986; Mojon, 2002). Cyclostratigraphic estimation of time according to Strasser & Hillgärtner (1998) and Strasser et al. (2004). Note that the time intervals between sequence boundaries are tied to the lithostratigraphy and not to the bio- and chronostratigraphy (e.g. the Privasensis subzone lasted only 500 kyr; the inferred 900-kyr time interval above sequence boundary Be 4 corresponds to a thin rock interval due to reduced sediment accumulation rate). For more discussion, refer to text. Tith.: Tithonian.
As explained above, it is suggested that one small-scale sequence lasted about 100 kyr and thus formed in tune with the short eccentricity cycle of the Earth’s orbit. The medium-scale sequences, commonly holding four small-scale sequences, then would represent the long eccentricity cycle of 400 kyr. Where five elementary sequences compose one small-scale sequence, they probably correspond to the precession cycle of about 20 kyr (Berger et al., 1989). However, autocyclical processes could also have led

| Medium-scale sequences | Small-scale sequences | Field aspect | Samples | Facies | Texture |
|------------------------|-----------------------|--------------|---------|--------|---------|
| Be 3                   |                       |              |         |        |         |
| 4                      | 2                     | 1            |         |        |         |
| 3                      | 2                     | 1            |         |        |         |
|                        | 5                     | 4            |         |        |         |
|                        | 3                     | 2            |         |        |         |
|                        | 1                     | 1            |         |        |         |
| 2                      |                       |              |         |        |         |
| 1                      |                       |              |         |        |         |
|                        | 5                     | 4            |         |        |         |
|                        | 3                     | 2            |         |        |         |
|                        | 2                     | 1            |         |        |         |

**Fig. 5.** Part of the Goldberg Formation in the Salève section (see also Fig. 6). Individual beds (separated by thin marly layers) are interpreted as elementary sequences. These are hierarchically stacked into small-scale and medium-scale sequences. The upper boundary of the medium-scale sequence is at the same time large-scale sequence boundary Be 3 (Fig. 4). Note the general thinning-up trend of beds up to Be 3, which indicates a general loss of accommodation space and corresponds to the late highstand of the previous large-scale sequence. Small-scale sequences 3 and 4 contain less than five elementary sequences, and the lithoclasts at sequence boundary Be 3 imply early cementation of sediment and subsequent reworking. Note also that it is useful to define a maximum-flooding zone because there is no unique surface that would have formed during maximum accommodation gain. For more discussion, refer to text (modified from Strasser & Hillgartner, 1998).
to the formation of such sequences, and their unequivocal attribution to an orbital cycle in many cases is not possible (Strasser, 1991; Hill et al., 2012). Even if some question marks remain, the cyclostratigraphic interpretation nevertheless offers a high-resolution time frame within which the sedimentological processes can be discussed.

DEPOSITIONAL ENVIRONMENTS

The predominant depositional environments on the shallow Jura platform have been reconstructed based on detailed microfacies analyses (Strasser, 1988; Pasquier, 1995; Hillgärtner, 1999; Tresch, 2007). These environments include:

- Emergent land characterized by calcrete crusts, root traces and circumgranular cracks. Black pebbles, commonly reworked in inter- or subtidal deposits, imply impregnation by organic matter that partly resulted from forest fires (Strasser & Davaud, 1983).

- Coastal lakes rich in charophytes (stems and oogons) and ostracods.

- Sabkhas as suggested by gypsum and anhydrite pseudomorphs as well as by brecciation that probably resulted from collapse after dissolution of evaporites.

- Tidal flats represented by mudstones bearing birdseye structures, desiccation polygons, and microbial laminations. Some of these stromatolitic laminae are dolomitized.

- Beaches composed of ooids, bioclasts and lithoclasts, and characterized by parallel lamination and keystone vugs. Locally, beachrock blocks are found.

- Shallow lagoons with peloids and oncoids. Abundance and diversity of benthic fauna and flora (foraminifera, gastropods, bivalves, brachiopods, echinoderms, serpulids, ostracods, dasycladalean algae) are variable, suggesting that some lagoons were restricted in terms of oxygenation and water energy, whereas others had open-marine conditions.

Fig. 6. Outcrop photograph of part of the Salève section, including sequence boundaries Be 3 and Be 4 (compare with Figs 4 and 5). Small-scale sequences below Be 3 are numbered as in Fig. 5. Note the slight dip of the stratification towards the right at the base of the Pierre-Châtel Formation (top of picture), which implies lateral migration of a large subtidal dune.
HIATAL SURFACES AND CONDENSED INTERVALS

In sequence-stratigraphic terminology, a hiatus is an interval of geological time that is not recorded in strata, and the corresponding surface along which this time is missing is an unconformity (Mitchum, 1977; Catuneanu et al., 2009). For this study, however, where also small-scale and short-term interruptions of sedimentation are discussed, the term ‘hiatal surface’ is preferred and defined as a discontinuity surface along which sedimentation has been interrupted, or which results from mechanical or chemical erosion of previously deposited sediment. Facies below and above the hiatal surface may be the same or different, and the time not recorded at these surfaces can vary from hours to millions of years.

In sequence stratigraphy, condensed sections are defined as thin stratigraphic intervals characterized by very low sediment accumulation rates that are mainly found in deep-water settings (Loutit et al., 1988; Catuneanu et al., 2009). Here, the term ‘condensed interval’ is used to describe a thin sediment body on the platform, in which much geologic time is represented. It may be composed of a single facies or be a composite of several thin sediment layers separated by closely spaced hiatal surfaces.

In the studied outcrops, beds are commonly separated by marly layers that are a few millimetres to a few centimetres thick. The beds can be followed over hundreds of metres where outcrop conditions allow (for example in the cliffs at Salève; Fig. 6). At the base of the beds the limestones may evolve gradually from the marls or else they are separated from them by sharp surfaces. The tops of the beds may pass into marls or again represent sharp surfaces. However, surfaces occur also within the beds where they are made visible through abrupt facies changes. These latter surfaces are of limited lateral extent. Table 1 summarizes the different types of surface recognized and their interpretation.

Simple surfaces of limited lateral extent

In oolitic and bioclastic grainstones, bed-parallel or oblique lamination can be observed that is created through differences in grain size. Erosion or cracking may follow softer lithologies and create visible surfaces (Fig. 7A). These structures are interpreted as reactivation surfaces having formed on laterally migrating dunes (Gonzalez & Eberli, 1997). Tidal influence is suggested by thin clay seams that occur locally and are interpreted as flaser bedding, and by bidirectionally dipping laminae (herringbone cross-bedding).

In limestone beds with a micritic matrix, irregular surfaces may occur, again rendered visible by facies contrasts. The best example is found in the Salève section (Fig. 7B) where an undulating surface separates bioturbated wackestone from dark-coloured floatstone with irregularly rounded clasts of the underlying facies. This feature is interpreted as the floor of a tidal channel. The relief of the surface was probably created by bioturbation and by
Tidal current scouring, which ripped up the cohesive sediment to form the soft pebbles. When current activity stopped, the bottom waters became stagnant to permit the partial preservation of organic matter, explaining the dark colouring. The channel fill then occurred in oxygenated conditions, and erosion of the channel flanks continued furnishing soft pebbles.

Around sequence boundary Be 1 in the Salève section, some surfaces marked by ripples and micrite-filled ripple troughs spread over a few metres but then disappear into thinly laminated, dolomitized beds (Fig. 7C and D). This configuration is interpreted as a tidal-flat environment with microbial mats and localized shallow depressions where currents and waves formed ripple marks (Bover-Arnal & Strasser, 2013). Thin marly flasers are also observed within these beds, again indicating tidal influence (Fig. 7D). Microbial lamination and desiccation into polygons are clearly visible below sequence boundary Be 2 at Salève (Fig. 7E). Chips break off along the lamina planes and furnish flat pebbles when reworked. The lamination is interpreted as being due to daily cycles of microbial growth on tidal flats (Hardie & Ginsburg, 1977).

Broken and tilted oolitic grainstone slabs occur at sequence boundary Be 2 at Salève (Fig. 7F). They are overlain by a badly sorted conglomerate, the components of which include the grainstone facies from the slabs below but also black pebbles and wacke- to packstones (Bover-Arnal & Strasser, 2013). Keystone vugs and parallel lamination in the oolites point to a beach environment, and the formation of the slabs is explained by early cementation to form beachrock, which was then broken by strong wave action. Locally, the broken oolite is covered by calcite, implying exposure in subaerial conditions before the deposition of the conglomerate.

In all cases described above, the lateral extent of the surfaces is limited to a few tens of centimetres to a few tens of metres. The processes that created these surfaces were of rather short duration: a few hours or days in the case of reactivation surfaces and laminites created by tidal activity, a few months to a few tens of years in the case of an abandoned tidal channel, and a few hundreds of years to cement carbonate beach sand and then break it into slabs (Halley & Harris, 1979; Strasser & Davaud, 1986). It has to be remembered, however, that the preserved sediments are snap shots of the geological history. For example, a tidal dune can migrate back and forth for hundreds of years before the current regime changes and the bedform is stabilized and recorded.

**Correlation of sections**

Outcrop conditions in the Jura Mountains do not allow walking out beds and surfaces over large distances. Therefore,

| Processes                              | Products                                      | Hiatal surfaces         | Lateral extent | Time lost in surfaces |
|----------------------------------------|-----------------------------------------------|-------------------------|----------------|-----------------------|
| Short-term                             |                                               |                         |                |                       |
| Waves and weak currents                | Ripples                                       | Reactivation surfaces   | Tens of centimetres | Hours                |
| Tides                                  | Tidal laminates, microbial mats, desiccation | Tidal-flat laminae      | Tens of centimetres to tens of metres | Hours to days |
| Strong tidal currents                  | Ooid and bioclastic shoals, grainstones, flaser bedding, herringbone cross-bedding, tidal channels | Reactivation surfaces, scour surfaces | Metres to tens of metres | Hours to tens of years |
| Storm waves, storm-induced currents    | Tempestites, HCS, SCS, rip-up clasts, flat-pebble conglomerates | Reactivation surfaces, scour surfaces | Metres to tens of metres | Hours to hundreds of years |
| Sediment bypass, sediment starvation   | Softgrounds, hardgrounds, bioturbation, biogenic encrustation, mineralization, reworked clasts | Firmground and hardground surfaces, maximum-flooding surfaces | Metres to tens of kilometres | Years to millions of years |
| Transgression                          | Lag deposits, rip-up clasts, deepening-up facies evolution | Transgressive surfaces, ravinement surfaces | Metres to tens of kilometres | Hundreds to thousands of years |
| Subaerial exposure                     | Karst, pedogenesis, soils, brecciation, calcrite, root traces, coal, freshwater diagenesis | Surfaces of maximum regression, sequence boundaries | Metres to tens of kilometres | Hundreds to millions of years |

Table 1. Comparison of short-term and long-term processes, associated sedimentary features, potentially produced hiatal surfaces and their lateral extent, and estimations of geological time lost in these surfaces.
Fig. 7. Outcrop photographs of short-term hiatal surfaces. (A) Reactivation surfaces in the subtidal dune at the base of the Pierre-Châtel Formation, dipping to the right (see also Fig. 6). These surfaces are enhanced by cracks formed along grain-size contrasts. Pen is 14 cm long. (B) Bottom of tidal channel in the Goldberg Formation, Saleve section. The lower part of the bed is strongly bioturbated (vertical traces filled with calcite cement). The bottom surface of the channel is irregular and wavy (to the right of pen). The lower part of the channel fill is dark-coloured and contains light-coloured soft pebbles reworked from the underlying sediment. Brownish-coloured soft pebbles occur in the upper part of the channel fill. The top of the bed appears in the upper left corner of the photograph. Visible part of pen is 6 cm long. (C) Lateral pinch-out of beds at sequence boundary Be 1 in the Saleve section. Some of the surfaces are laterally continuous, others disappear. Arrow points to ripple trough shown in D. Hammer is 33 cm long. (D) Detail of C, showing a ripple trough filled with brownish micrite. Thin marly seams (flasers) are visible below and above (arrows) but are not continuous laterally. (E) Desiccation polygons below sequence boundary Be 2 in the Saleve section. Note chips breaking off along tidal laminae. Visible part of red pen is 3 cm long. (F) Broken beachrock slabs overlain by conglomerate at sequence boundary Be 2, Saleve section. Note the dark-grey or black pebbles. Visible part of pen is 10 cm long.
measured sections have first been interpreted as described in the Methods chapter and then correlated. Figure 8 shows such a correlation between four sections representing the shallow-water platform carbonates of the Pierre-Châtel Formation. A correlation with the hemipelagic facies of the Montclus section in the Vocontian Basin (south of France) is added, where the ammonite biostratigraphy is well established and allows the age of the shallow-water sections to be constrained (Fig. 4; Strasser et al., 1998). The limestone-marl couplets at Montclus are interpreted to have formed in tune with the 20 kyr orbital precession cycle. These couplets commonly group into bundles of five, indicative of the 100 kyr short eccentricity cycle. The bundles correspond to the small-scale sequences identified in the platform sections. Thus, major sequence boundaries identified and dated by Hardenbol et al. (1998) in many European basins can be identified and correlated from deep-water to shallow-water sections.

In Fig. 8, sequence boundaries (SB) Be 4 and Be 5 are indicated. The base of the Pierre-Châtel Formation corresponds to the transgressive surface (TS) above SB Be 4. At Montclus, however, transgressive surfaces (TS 1 and TS 2 in Fig. 8) are suggested by rapid changes from more limestone-dominated couplets to more marl-dominated ones, implying that the physical expression of transgression did not occur at the same time in the hemipelagic basin as on the platform (Pasquier & Strasser, 1997). At Montclus, the maximum flooding is interpreted to be represented by the most marly interval of this outcrop, which corresponds to marly and strongly bioturbated facies in the platform sections. Sequence boundary Be 5 at Montclus can be placed either at the base of a thick limestone bed (bundle 19) or else at the base of the slumps (23-3 to 25-2 m); consequently, a sequence-boundary zone is proposed that covers this interval of uncertainty. The nine hemipelagic bundles at Montclus (representing 900 kyr) above SB Be 4 are interpreted to be condensed into half a metre of sediment at Salève. At Montclus, 1-8 to 1-9 Myr are recorded between SB Be 4 and SB Be 5, whereas at Chapeau du Gendarme and Crêt de l’Anneau the record of this large-scale sequence ends with small-scale sequence 17, and at Rusel with number...
15. At Salève, a gap in the outcrop does not allow the placement of SB Be 5.

In the following, four sequence-boundary zones and one maximum-flooding zone will be discussed in detail.

**Complex intervals of wide lateral extent**

The irregular beds with lateral pinch-outs in the lower part of the Salève section (Fig. 7C) can be followed laterally over a few tens of metres. They show a dramatic facies change from high-energy, subtidal, marine grainstones in the massive bed below passing upwards to inter- to supratidal dolomitized microbial laminites including black pebbles and wood fragments, followed by lacustrine wackestones (Bover-Arnal & Strasser, 2013). The charophyte-ostracode assemblage M 1a suggest an early Berriasian age, and the sequence- and cyclostratigraphic interpretation of Strasser & Hillgärtner (1998) suggests that this interval corresponds to sequence boundary Be 1 (Fig. 4). Unfortunately, outcrop conditions in the other studied sections do not allow this boundary to be confidently correlated over large distances. However, even at the decametre scale, the complexity of this sequence-boundary zone is evident, with some surfaces and beds being laterally continuous, others having a limited extent.

The boundary between the Goldberg and the Pierre-Châtel formations is very well-developed and can be followed over the entire study area. In the Salève section, this boundary corresponds to a complex interval comprising peritidal facies with microbial laminites and desiccation polygons followed by broken beds of various marine, peritidal and lacustrine facies (Fig. 9A; Bover-Arnal & Strasser, 2013). Some clasts are blackened, suggesting supratidal conditions (Strasser & Davaud, 1983). In other sections, similar breaking and reworking of pre-existing limestones is found (Fig. 9B). This interval is attributed to sequence boundary Be 4, well dated by ammonites (Fig. 4; Clavel et al., 1986). The transgression leading to the open-marine, high-energy facies at the base of the Pierre-Châtel Formation occurs in two pulses in the Salève section (Fig. 9A), whereas at Rusel, it is more gradual (Fig. 9B).

At Crét de l’Anneau, the top of small-scale sequence 12 is developed as an undulating, reddish surface bearing dinosaur tracks (Pasquier, 1995). It is covered by a bioturbated wacke- to packstone that pinches out laterally (Figs 8 and 9C, D, E). Assuming that the correlation presented in Fig. 8 is correct, this same surface terminating sequence 12 shows karst features due to subaerial exposure at Rusel (Fig. 9F) and birdseye structures of a tidal-flat environment at Chapeau du Gendarme. At Salève, its precise correlation is uncertain as the whole section is composed of subtidal facies (Fig. 8). The wedge of bioturbated limestone at Crét de l’Anneau develops laterally into a full-fledged small-scale sequence (number 13 in Fig. 8). In parts of the outcrop at Crét de l’Anneau, only a thin marly layer separates small-scale sequences 12 and 14; sequence 13 is missing completely (Fig. 9 E). At Rusel, the top of sequence 12 not only contains karst features and yellow-red staining but also borings (Fig. 9F), indicating that the surface was hardened before marine flooding (Tresch, 2007).

The best candidate for the maximum-flooding zone of large-scale sequence Be 4 is the strongly bioturbated interval in small-scale sequence 16 that is commonly expressed by a softer lithology in the field (Figs 8 and 10A, B). Concentration of bioturbation is here interpreted as being a sign of condensation: at low sediment accumulation rates, the same sediment is reworked by many generations of burrowing organisms. The low accumulation rate may be the result of low carbonate production in deeper and/or more turbid water where photosynthesis of the carbonate-producing organisms (e.g. green algae) is reduced, and/or of removal of sediment by currents. The increased marl content reflects water depth below wave base where the clay minerals were not winnowed away (Strasser & Hillgärtner, 1998). At Crét de l’Anneau, deepening of the environment must have occurred rapidly because the top of small-scale sequence 14 still shows karst features and the overlying bed pinches out laterally, probably a sign of channelling (Fig. 10B). At Chapeau du Gendarme, the top of small-scale sequence 15 (Fig. 10C) is interpreted as a small-scale sequence boundary because the bed below contains circumgranular cracks implying pedogenesis (Fig. 8). The perforated hardground capping the thin bed above formed through sediment starvation, announcing the maximum flooding in the overlying sequence 16. In the hemipelagic section of Montclus, the most marly part is found in sequence 15. This discrepancy can be explained by the fact that not all places on the platform and in the basin reacted in the same way and at the same time to a rapid sea-level rise (Strasser et al., 2000). The position of this maximum-flooding zone in the Be 4 sequence is strongly asymmetric (Fig. 8). The same asymmetry is observed in other European basins (Hardenbol et al., 1998), suggesting that the associated long-term sea-level change also was asymmetric. At Rusel, this maximum-flooding interval is not recorded. Nevertheless, the relatively thick small-scale sequence 15 there (Fig. 8) indicates that accommodation was high, which might have been due to an increased rate of sea-level rise (but see also the discussion on synsedimentary tectonics below).
Fig. 9. Outcrop photographs of sequence boundaries and transgressive surfaces with a wide lateral extent (for numbering of sequences and correlations see Fig. 8). (A) Sequence-boundary zone Be 4 at Salève, at the top of the Goldberg Formation. Note bed limit above hammer to the left, while to the right of photograph the beds are completely dismantled. A laterally continuous bed delimited by two sharp transgressive surfaces at the base of the Pierre-Châtel Formation covers the sequence-boundary zone. Hammer is 33 cm long. (B) Sequence-boundary zone Be 4 at Rusel. The marly Goldberg Formation below and the yellowish, massive Pierre-Châtel Formation above are separated by dismantled limestone beds, and no clear transgressive surface is developed. (C) Sharp reddish surface at the top of small-scale sequence 12 in the Crêt de l’Anneau section. Irregular depressions in the surface are possibly due to dinosaur tracks. Bioturbated wacke- to packstones are onlapping on this surface from the right (arrow at contact). The bioturbated facies constitutes small-scale sequence 13. (D) Detail of bioturbated facies (upper half of photograph) overlying the sharp surface. Pen is 14 cm long. (E) Only a thin marly layer separates the sharp surface (interpreted as sequence boundary) at the top of small-scale sequence 12 and the transgressive surface forming the base of small-scale sequence 14. Sequence 13 is not recorded. (F) Red-stained surface on top of small-scale sequence 12 in the Rusel section. Circular structures are borings (probably Gastrochaenolites), suggesting marine flooding after subaerial exposure.

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Fig. 10. Outcrop photographs of maximum-flooding zone and surface, and sequence boundaries with a wide lateral extent (for numbering of sequences and correlations see Fig. 8). (A) Nodular, marly, strongly bioturbated interval in the Crét de l’Anneau section, composing the upper half of small-scale sequence 15 and sequence 16. The limit between these two sequences is not well defined but suggested by a softer, more marly passage that is laterally consistent (dashed white line). This interval is interpreted as the maximum-flooding zone of large-scale sequence Be 4. (B) Irregular, reddish surface at top of small-scale sequence 14 in the Crét de l’Anneau section. Note lateral pinch-out of the overlying bed (arrow). Hammer is 33 cm long. (C) Sharp, bioperforated hardground surface (at the base of hammer) on top of the thin bed below the base of sequence 16 at Chapeau du Gendarme. It is interpreted as a first manifestation of the large-scale maximum flooding. (D) Interval of thin beds at Chapeau du Gendarme, with a hardground on top of the thick bed below (small-scale sequence 17). This hardground surface is interpreted as sequence boundary Be 5. The surface below the hardground pinches out laterally and possibly was a channel floor (arrow). (E) Yellow-reddish, irregular surface (at base of hammer) interpreted as large-scale sequence-boundary Be 5 in the Crét de l’Anneau section. (F) Same surface as in E, showing the deep penetration of the iron staining (face of bed in lower left corner of picture).
Sequence-boundary zone Be 5 is placed below a slumped interval in the hemipelagic Montclus section and well dated by ammonites (Figs 4 and 8). In the platform sections of Rusel, Crêt de l’Anneau, and Chapeau du Gendarme, SB Be 5 separates the Pierre-Châtel Formation from the Vions Formation and is developed as a karstified surface (unfortunately, it is covered in the Salève section).

At Chapeau du Gendarme, this surface caps lagoonal sediments with an oblique stratification suggesting a tidal channel (Figs 8 and 10D). It is overlain by thin, bioturbated beds with marine fauna (echinoderms), implying rapid flooding after the subaerial exposure. At Rusel and Crêt de l’Anneau, the surface displays intense red-yellow iron staining that penetrates a few tens of centimetres into the underlying limestone (Fig. 10E and F). Stable-isotope analyses carried out in these two sections (Pasquier, 1995) show a positive shift in δ18O just below the surface, while δ13C has a negative shift (Fig. 11). This is explained by subaerial exposure that increased evaporation (concentrating the heavy oxygen isotopes), while soil gas led to concentration of light carbon in the early-diagenetic fluids. This particular composition was then preserved through mineralogic stabilization and cementation of the sediments (Joachimski, 1994). Such a signature is not present at the surface on top of small-scale sequence 14 (Fig. 11), probably due to a shorter subaerial exposure time there than at the surface attributed to sequence-boundary Be 5.

From the examples presented above it is evident that surfaces of wide lateral extent are rarely simple surfaces but complex intervals that change their aspect from one outcrop to another. In many cases they are amalgamations or stacks of closely spaced simple surfaces, each of which has only local extent (e.g. Figs 7C and 10D). Correlation based on physical expression alone therefore can be difficult, and it is rather the position in the stratigraphic column that demonstrates their relationship. Sequence boundaries of basin-wide scale are commonly composites of several types of simple surface that, however, all express loss of accommodation space on the platform. Transgressive surfaces in the studied sections commonly are sharp, whereas maximum-flooding conditions can be expressed by a single hardground but also by an interval with dense bioturbation and increased clay content.

**DISCUSSION**

**From the living platform to the depositional record**

On modern, shallow carbonate platforms it is easy to directly observe the short-term processes that lead to the formation of distinct surfaces: waves and tidal currents erode, transport, and re-deposit carbonate grains and mud, and strong tidal or storm-induced currents scour channels into previously deposited sediment (Table 1). These processes lead to short-term hiatuses in which a few hours (in the case of ripple migration or tidal-flat lamination) to a few hundreds or thousands of years (in
the case of deep scour surfaces) of sedimentation history are lost. On active ooid or bioclastic dunes, the sediment is mobile and the hiatuses are represented by reactivation surfaces (McCabe & Jones, 1977; Gonzalez & Eberli, 1997). Incipient hardgrounds having formed by microbial binding and rapid cementation on the lagoon floor or on inactive dunes can be reworked to produce lithoclasts that are incorporated into the overlying sediment package (Hillgärtner et al., 2002). Erosion of cohesive and commonly microbially colonized carbonate mud on tidal flats and in low-energy lagoons produces soft pebbles that are again incorporated into the overlying deposit. The corresponding hiatus surfaces are cutting irregularly into the underlying soft sediment. On the upper parts of tidal flats, non-deposition occurs at low tide during a few hours. Strong tides or storms ripping up the microbially stabilized tidal laminae produce flat pebbles (Hardie & Ginsburg, 1977).

Bioturbation leads to homogenisation and time-averaging of the uppermost sediment layer (Shinn, 1968; Flessa et al., 1993). Thus, hydrodynamic sedimentary structures and hiatus surfaces may be destroyed and detailed information about a few years to hundreds of years of sedimentary history is lost. Sediment bypass and sediment starvation lead to the formation of softgrounds that turn into firmgrounds and hardgrounds when stabilized by biogenic and/or mineral encrustations and by early diagenesis (Christ et al., 2015). Years to hundreds of years of sedimentary history may thus be condensed into a thin interval or a single surface (Table 1). With continued sediment starvation, hardgrounds may also represent millions of years but in this case represent an amalgamation of multiple processes such as repeated phases of flooding and subaerial emergence (Rameil et al., 2012).

Hiatal surfaces produced by short-term processes and covering hours to a few hundreds of years commonly are of limited lateral extent (Table 1). This is due to the facies mosaics typical of shallow carbonate platforms where different depositional environments are closely juxtaposed (Rankey, 2002; Rankey & Reeder, 2010). For example, a tidal channel produces a scoured surface that is only a few metres wide, and an individual reactivation surface in an ooid shoal covers at the most a few tens of square-metres, corresponding to the size of the inter- or subtidal dunes forming the shoal. The same holds for short-term condensation, for example expressed by intense bioturbation over a few hundreds of square-metres in a quiet and sediment-starved depression behind the shoals.

With the knowledge of short-term processes and products observed on modern carbonate platforms, it is relatively straightforward to interpret the features seen in the sedimentary record. Fossil examples of such hiatus surfaces with limited lateral extent have been presented in a previous section (Fig. 7).

When long-term (i.e. involving thousands to hundreds of thousands to millions of years) rises and falls of relative sea-level flood or expose a carbonate platform, the above-mentioned short-term processes interact and produce complex amalgamations and stacks of hiatus surfaces or condensed intervals (examples in Figs 9 and 10). These surfaces and intervals can be interpreted in terms of sequence stratigraphy (Vail et al., 1991; Catuneanu et al., 2009).

Sequence boundaries on a shallow carbonate platform form when sediment has completely filled the available space (created by eustatic sea-level and subsidence) and reached intertidal to supratidal conditions (at the top of shallowing-upward highstand deposits). If eustatic sea-level falls below the previously accumulated sediment surface or if there is tectonic uplift, erosion of mobile sediment, cementation in a fresh-water lens, karstification and pedogenesis occur. Depending on the amplitude of relative sea-level fall and on the time of subaerial exposure, important amounts of previously deposited sediment may be lost through mechanical erosion and/or chemical dissolution. In the sedimentary record, this can lead to superposition of subaerial exposure surfaces directly on subtidal facies, or to the erosion of entire previously deposited sequences. In the case of a deep lagoon and a low-amplitude sea-level fall, facies may stay subtidal throughout the sequence. Sequence boundaries there may be expressed only indirectly by input of siliciclastics eroded from the hinterland (Osleger, 1991; Strasser & Hillgärtner, 1998).

Transgressive surfaces form when relative sea-level rise leads to flooding of the platform. On tidal flats, accommodation is created but quickly filled by flourishing microbial mats until sea-level rise outpaces the accumulation potential. On beaches, storm surges will create ravinement surfaces and push back the beach ridges until the system is drowned by continued rising sea-level (Danselaar, 1989). On previously subaerially exposed surfaces, the flooding will pick up lithoclasts derived from the underlying cemented sequence, pedogenic elements and plant fragments, and incorporate them at the base of the deepening-up transgressive deposits (Strasser & Davaud, 1983). A certain lag time will pass before the organic carbonate factory starts up again and sediment production keeps up with rising sea-level (_tipper, 1997; Kemp & Sadler, 2014).

Maximum-flooding conditions on a shallow carbonate platform are indicated by the turnaround from a deepening-up to a shallowing-up facies evolution. This change may be gradual and produce an interval of sediment,
without the development of a specific surface. However, when the fastest relative sea-level rise is accompanied by sediment starvation, a strongly bioturbated surface may form in lagoons, which likely will develop into a hardground (Reolid et al., 2014). In the studied sections, the condensed intervals are enriched in clay minerals, suggesting that terrigenous input possibly hampered the carbonate-producing organisms and thus contributed to a low sedimentation rate.

Figure 12 schematically illustrates the formation of two sequence boundaries, a transgressive surface, and a maximum-flooding surface on a shallow carbonate platform. It is assumed that one symmetrical sea-level cycle is superimposed on a longer term rising trend that, together with subsidence, creates accommodation space. When sea-level first drops below the pre-existing sediment surface, the soft sediment is eroded. Cementation in the fresh-water lens has in the meantime stabilized the sediment, and karstification occurs. This karstified surface will be recognized as the first sequence boundary. When sea-level rise leads to flooding of the karst surface (placing the transgressive surface directly over the sequence boundary), lithoclasts will be reworked, forming a lag deposit. Carbonate production and sediment accumulation will pick up after a lag time. Water depth at first is ideal to produce ooid dunes, shifted by tidal currents. Further deepening leads to lagoonal facies, which keeps up with rising sea-level. If a storm event happens, the sediment surface may be lowered through erosion of the lagoonal sediment. Sediment starvation may lead to a hardground (that will be interpreted as the maximum-flooding surface) and a strongly bioturbated condensed interval, corresponding to the deepest water. With diminishing water depth, carbonate production picks up again and sediment starts filling the lagoon, until falling sea-level forces erosion and a second phase of karstification (the second sequence boundary). This hypothetical scenario describes the evolution of only one point on the platform and may (at the same time) be entirely different at other points of the same platform, depending on the substrate morphology and the lateral facies distribution. Also, rates of carbonate production and sediment accumulation vary strongly throughout this one cycle.

In a two-dimensional vision of the platform (Fig. 13), it is schematically illustrated how areas of sediment accumulation and areas of concomitant erosion and sediment reworking are juxtaposed. Sea-level fluctuations with two superimposed frequencies control the gain or loss of accommodation space. Depending on the initial platform morphology and the sediment produced in lakes, on beaches, on shoals, in lagoons or on reefs, the distribution of facies and of hiatal surfaces will vary dramatically. Accordingly, the resulting sedimentary records will differ depending on their position on the platform. It is also demonstrated that the sequence boundaries and transgressive surfaces as identified according to facies changes in each record will not necessarily have formed during the same sea-level cycle, or are amalgamations when the record of one or more cycles is missing. For example, record D represents all five sea-level cycles, while record E shows only two. In the position of record C, a beach formed during cycle 1. Its preservation is possible because of a first transgressive pulse above the sequence boundary. A second transgressive surface then leads to the installation of ooid dunes. In record D, the lagoon was filled in during cycle 1. The presence of lithoclasts and plant fragments in the overlying bed (2) implies the presence of a sequence boundary (position b on the sea-level curve), directly overlain by a second transgressive surface. Lowstand deposits are missing because of the low accommodation space. The maximum-flooding interval always appears within cycle 4 when much accommodation is
Fig. 13. Hypothetical evolution of a platform transect during five high-frequency sea-level cycles. Note that the completeness of the sedimentary record changes strongly depending on the position on the platform. Cross-sections a to e reflect the platform morphologies and the facies distributions at the time steps a to e indicated on the sea-level curve. For explanation, refer to text.
created by the superposition of a short-term sea-level rise on the high of the lower frequency curve (position d). However, in record C there is no facies change that would allow identification of this maximum flooding. Note that on shoals and on beaches erosion and accumulation may occur at the same time as sediment is reworked during any time of a sea-level cycle. It is clear that the picture is even more complex if a real-world, three-dimensional platform is considered. In addition, synsedimentary tectonics may modify the accommodation space at irregular intervals and differentially from one place on the platform to another.

Not only cyclical sea-level changes but also random processes influence the sedimentation on shallow carbonate platforms. For example, lateral migration of sediment bodies may create shallowing-up facies successions at constant sea-level (Pratt & James, 1986; Satterley, 1996) that look similar to the ones produced by sea-level fluctuations (Strasser, 1991). Storm events may create erosion surfaces at any time, independent of a sea-level cycle. Burgess & Wright (2003) and Burgess (2006) showed by forward modelling that changes in carbonate production rate and changes in sediment transport direction may produce repetitive strata, independent of sea-level changes. Nevertheless, the hierarchical stacking of depositional sequences (Fig. 5) and the relatively good fit with the time frame given by Hardenbol et al. (1998) (Fig. 4) suggest that the periodicities of the orbital cycles are reproduced in the studied sections. In Berriasian times, continental ice probably was present (Fairbridge, 1976; Frakes et al., 1992; Eyles, 1993) but ice volumes were small and the glacio-eustatic fluctuations (resulting from orbitally controlled insolation changes) of low amplitude. However, these same insolation changes also caused thermal expansion and retraction of the uppermost layer of ocean water (Gornitz et al., 1982; Wigley & Raper, 1987), thermally induced volume changes in deep-water circulations (Schulz & Schäfer-Neth, 1998), and/or water retention and release in lakes and aquifers (Jacobs & Sahagian, 1993). These processes contributed to high-frequency, low-amplitude sea-level changes (Plint et al., 1992; Conrad, 2013). Consequently, it is assumed that the cyclical insolation changes translated rather directly into sea-level fluctuations that were therefore more or less symmetrical (in contrast to the asymmetric glacio-eustatic fluctuations during ice-house times when there was slow waxing and rapid waning of polar ice caps). Based on the Chapeau du Gendarme section, Strasser et al. (2004) reconstructed amplitudes of a few metres for a sea-level cycle related to the orbital short eccentricity (100 kyr) cycle. On a shallow platform, even such low amplitudes will lead to significant facies changes. In addition, orbitally controlled climate changes may influence rainfall in the hinterland and thus modulate siliciclastic and nutrient input onto the platform. Not only orbitally induced processes in the Milankovitch frequency band but also millennial-scale cyclical processes may create laterally consistent bedding planes. For example, Zülhike et al. (2003) identified high-frequency cyclicity producing shallowing-upward sequences in the Triassic Latemar platform, and Tucker et al. (2009) explained mid-Carboniferous cyclothems by high-frequency arid-humid climate changes that were superimposed on climate and sea-level changes induced by the orbital precession cycle. These high-frequency climate changes were possibly controlled by variations in solar activity (Elrick & Hinnov, 2007).

Consequently, the final sedimentary record results from a combination of allocyclic (externally controlled) and random processes. Whereas allocyclic processes such as orbitally controlled climate and sea-level changes, or millennial-scale climate changes, influence the entire platform, random processes are of limited lateral extent. In the studied sections, it is assumed that the formation of small- and medium-scale sequences that can be correlated across the entire Jura platform was at least partly controlled by orbitally induced eustatic sea-level cycles. In the case of some elementary sequences, however, random processes dominated and blurred the low-amplitude sea-level signal. The lateral variability in the thickness of sequences (Fig. 8) is attributed not only to the variable potential of sediment accumulation in the different environments (Fig. 13) but also to synsedimentary tectonics that structured the Jura platform into high and low areas (Hillegartner, 1999). This structure changed through time, thus modifying the substrate morphology and the accommodation at irregular intervals.

Duration of hiatuses and condensation

Estimation of time lost in hiatal surfaces or condensed in sequence-boundary or maximum-flooding zones is not an easy task. Time lost means that parts of the sedimentary history are not recorded at all, and time condensed means that many events of a long history are amalgamated and overprint each other in a thin layer of sediment that may be difficult to disentangle.

For simple surfaces, comparisons with present-day processes are helpful. For example, reactivation surfaces in tidally influenced ooid dunes are controlled by switches in current direction and current strength, and time lost at such surfaces is in the order of a few hours (Hine, 1977; Gonzalez & Eberli, 1997). However, reactivation surfaces may also be induced by spring tides (twice-monthly), equinoxial tides (twice-yearly), or episodic storms, which then increases the time span manifold. Estimation of the geological time actually represented by a tidal dune is
Surfaces within the massive bed imply dune migration. Lateral correlation suggests that the top of sequence 12 exhibits iron staining and dinosaur tracks (Fig. 9C). The top of sequence 12 is interpreted as a transgressive surface, the cross-stratified oolitic and bioclastic grainstone. Its base is illustrated such a case (Fig. 14A). It is a massive bed of sandstone composed of the same facies are present, suggesting that incipient hardgrounds had formed and were reworked. Consequently, time lost in such surfaces could have been a few months to a few tens of years. The conclusion is that almost all of the 100 kyr available between the two sequence boundaries is lost or condensed in the marly layer below the bed and in the reddish surface on top of it. Additional, short time intervals are lost along the reactivation surfaces. It is impossible to say at which time within the 100-kyr cycle the dune was deposited, and if (and how much) other sediment was deposited but then eroded during this time interval.

In low-energy settings, simple surfaces form for example when tidal channels are cut into tidal flats. In the Bahamas, tidal channels stay active over tens of years with only little lateral migration (Rankey & Morgan, 2002), and their bottom will thus form a hiatal surface where this time interval is not recorded. Time is lost also on supratidal flats where sediment accumulates only during storm washovers, although algal-microbial growth may contribute to some accretion (Hardie & Ginsburg, 1977). In shallow lagoons, time may be condensed when sediment production and accumulation is low, which will result in strongly bioturbated intervals and hardgrounds.

An example of a low-energy setting is shown in Fig. 14B. Small-scale sequence 10 of the Chapeau du Gendarme section is composed of three beds of wackestone-packstones. Echinoderms and benthic foraminifera indicate a marine environment but charophytes were washed in from nearby coastal lakes, or reworked from now eroded lacustrine facies. Birdseye structures imply periodic inter-to supratidal exposure. The lower boundary of this sequence is not clearly developed but interpreted to sit in the thick marly bed that carries charophytes and black pebbles. The upper boundary is placed in the marly, charophyte-bearing layer separating it from sequence 11. Lateral correlation suggests that the time available to form small-scale sequence 10 was 100 kyr (Fig. 8). The base of the lowermost limestone bed is seen as an important transgressive surface that can be correlated over the entire study area (base of the Pierre-Châtel Formation; Fig. 8). According to the cyclostratigraphic scheme discussed above and illustrated in Fig. 5, it is assumed that each limestone bed represents 20 kyr. The beds are separated by thin marly layers, implying a change in environmental conditions and input of clay minerals (Strasser & Hillgartner, 1998). At the bottom of the tidal channel (top of the first bed), a few tens to a few hundreds of years were possibly lost. There is no evidence for other hiatal surfaces, but the slightly more marly and bioturbated levels within the

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**Fig. 14.** Examples of unequal time distribution in a sequence formed in a high-energy setting (A: small-scale sequence 12 at Crêt de l’Anneau) and in a sequence reflecting a low-energy context (B: small-scale sequence 10 of the Chapeau du Gendarme section). Symbols and abbreviations as in Fig. 8. For discussion, refer to text.
beds may express some condensation. The partings between the beds are interpreted as sequence boundaries directly followed by transgressive surfaces of the elementary sequences. At transgressive surfaces, a lag time may be lost, which has been estimated to be some 3000 years for shallow carbonate platforms (Kemp & Sadler, 2014). However, considering the various facies and laterally changing environments in shallow carbonate systems, this lag time may strongly vary from one position on the platform to another. Today, tropical lagoonal carbonate sediments accumulate with rates of a few tenths of a millimetre to a few millimetres per year (Enos, 1991; Strasser & Samankassou, 2003). Each of the three beds is about 60 cm thick, but decompacted it would have corresponded to about 1-2 m of sediment (Goldhammer, 1997). At an assumed average accumulation rate of 1 mm yr\(^{-1}\), it would have taken about 1200 years to make one bed. Consequently, it can be concluded that most of the 20 kyr available for the formation of one bed must be contained in the partings at the bed limits. As only three beds are recorded but the entire small-scale sequence lasted 100 kyr, additional time was lost and/or condensed in the marly intervals below and above the three limestone beds (Fig. 14B).

Complex, composite surfaces that correspond to sequence-boundary zones are in fact an amalgamation of different types of simple hiatal surfaces. As a result of reduced accommodation space on the shallow platform, processes creating hiatal surfaces are common: reactivation of dunes and beaches, ravinement surfaces related to storms, cutting of tidal channels, and subaerial exposure. As long as the sediment is not consolidated, erosion may partly or completely remove previously deposited material. Once the sediment is cemented, erosion and reworking will produce clasts that are incorporated into the overlying deposit. Cementation of beachrock and coastal carbonate dunes can happen within a few hundreds to thousands of years (Halley & Harris, 1979; Strasser & Davaud, 1986). Estimation of time distribution in such complex intervals is difficult. Sequence-boundary zone Be 1 (Fig. 7C) is characterized by well-defined beds that, however, pinch-out laterally. These beds may have formed under the influence of low-amplitude sea-level fluctuations driven by the 20 kyr precession cycle, or else they may be related to autocyclical, random processes of unknown duration (Strasser, 1991). As there is no lateral continuity of these beds, it is difficult to evaluate the time lost in the surfaces. In the case of sequence-boundary zone Be 4, about 900 kyr are represented by just half a metre of sediment (Figs 8 and 9A). Even by analysing all relict beds and the reworked clasts it is not possible to say at which time within these 900 kyr which facies was formed.

Sequence boundary Be 5 on the platform is developed as a sharp, unique surface (Figs 8 and 10D, E) although it certainly contains a complex history of deposition, erosion, and subaerial exposure (the latter evidenced by C and O isotopes; Fig. 11). According to the correlation in Fig. 8, 100 to 300 kyr are missing depending on the palaeogeographical position of the outcrops, which is explained by a tectonic tilt of the platform (Hillgärtner, 1999). A more local tectonic event may have been responsible for the thin or even absent small-scale sequence 13 in the Crêt de l’Anneau section (Figs 8 and 9C, E).

Maximum-flooding zones in the studied shallow-water outcrops commonly display intense bioturbation (Fig. 10A). The only way to estimate the time represented within these zones is lateral correlation and a cyclostratigraphic analysis. In the case of sequence 16 of the Crêt de l’Anneau section, 100 kyr are condensed in about 1 m of sediment. There is no evidence of hardgrounds or other hiatal surfaces, implying that it was a low sedimentation rate that led to this concentration of bioturbation. Also sequence 13 at Crêt de l’Anneau is strongly bioturbated. In addition to low sediment accumulation rate, it was reduced accommodation that induced the very thin deposit and the lateral pinch-out (Figs 8 and 9C).

**Sedimentation rates**

For the purpose of this paper, three types of ‘sedimentation rate’ are distinguished:

1 ‘Carbonate production rate’ describes the sediment furnished by the carbonate-producing organisms (measured in kg m\(^{-2}\) yr\(^{-1}\) for weight or mm per year for vertical accretion in modern environments; Enos, 1991).

2 ‘Sediment accumulation rate’ refers to the sediment that is accumulated in one spot on the platform (measured in mm per year in modern systems or estimated in ancient rocks). Sediment accumulation rate may be equal to carbonate production (accretion) rate if there is no sediment transport away from or into the production area, and if there are no hiatuses. To compare production and accumulation rates, the sedimentary record must be decompacted (Strasser & Samankassou, 2003). Sediment accumulation rates in ancient rocks should be estimated over short time intervals in order to better compare them with modern rates and interpret the sedimentary processes and facies evolution on the platform.

3 ‘Preservation rate’ is calculated by dividing the thickness of the rock record by the time interval it covers (mm yr\(^{-1}\) to m Myr\(^{-1}\)). Preservation rate is generally...
less than accumulation rate because hiatuses occur in the rock record and the originally accumulated sediment has been compacted. Preservation rates are useful to describe the long-term evolution of a platform, including long-term changes in sea-level and subsidence. In many publications, ‘sediment accumulation rate’ is calculated based on the preserved rock record representing millions of years and would thus correspond to the ‘preservation rate’ introduced here (Sadler, 1981, 1999; Bosscher & Schlager, 1993; Schlager, 1999).

Considering the abundance of hiatuses and condensations in shallow-water carbonate sediments, it is of course difficult to calculate sediment accumulation rates. Bosscher & Schlager (1993) compiled published accumulation rates and estimated a maximum of 200 m Myr\(^{-1}\) for Phanerozoic carbonate platforms, but they did not consider compaction or the presence of hiatuses. Sadler (1981, 1999) and Schlager (1999) discussed the completeness of the stratigraphic record and the scaling of sedimentation rates: apparently, sediment accumulation rates decrease as the time interval over which they are averaged increases. Schlager et al. (1998) stated that this is due to ‘the fact that sedimentation is an episodic process and that the sediment record is riddled with hiatuses on all scales’. If ancient sediment accumulation rates want to be estimated and compared to the ones in modern systems, it is more appropriate to consider only short time intervals in which hialtal surfaces and condensed intervals are absent or at least minimal. Furthermore, the sedimentary record must be decompacted according to facies (Goldhammer, 1997; Strasser & Samankassou, 2003). In the examples of the Jura Mountains discussed here, where overburden has never been more than about 1 km (Trümpy, 1980) and tectonic compaction can be excluded, one metre of sedimentary record of an ooid shoal would have corresponded to 1-1.2 m of mobile ooid sand. However, as discussed above, this represents only the final accumulation and not the production potential of an ooid shoal where most of the grains are washed back and forth and often are also exported to deeper basins (Rankey & Reeder, 2010). A bed of lagoonal wackestone measuring 1 m today would have corresponded to some 2 m of soft sediment. If such a bed is interpreted to have formed under the control of an orbital precession cycle of 20 kyr, then a very low sediment accumulation rate is calculated (0-1 mm yr\(^{-1}\)), which is ten times less than values seen in modern lagoons (average of 1 mm yr\(^{-1}\); Enos, 1991). Assuming that the ecological conditions for the carbonate-producing organisms in the Berriasian were similar to the ones today (and thus the accumulation rates are comparable), it has to be postulated that much time is lost at the bedding planes.

Furthermore, carbonate production and accumulation rates may vary significantly throughout the time interval during which the bed was accumulated, and also laterally across the platform (Figs 12 and 13; see also Strasser et al., 2012, for an Oxfordian case study).

These two examples illustrate that, to calculate a sedimentation rate, it is misleading to just divide the thickness of a section by the number of years it represents. Such a calculation informs at the most about an average preservation rate. It is much more informative to calculate the rates of specific intervals (decompacted according to facies) in the sedimentary record where it can be assumed that sedimentation was continuous. Even so, it will be no more than an average sediment accumulation rate because lateral changes of facies and sediment thickness can strongly vary on a shallow platform. Furthermore, sediment accumulation rate may be much less than carbonate production rate because much of the material produced in one site may be shifted across the platform by currents or exported to a deeper basin (e.g. Schlager, 1991; Milliman et al., 1993). In many cases, on a shallow platform, accumulation is limited not by carbonate production but by accommodation once the available space has been filled by sediment.

To illustrate this issue, the section of Chapeau du Gendarme is analysed in detail (Table 2). Facies are decompacted assuming that marls and mudstones have been more compacted than textures containing grains such as wackestones, packstones, and grainstones (Strasser & Samankassou, 2003). Then, modern sedimentation rates (averaged from Enos, 1991) are used to calculate the time theoretically needed to accumulate the decompacted facies, and this time is compared with the total time span between sequence boundaries Be 4 and Be 5. While Be 4 has not been identified at Chapeau du Gendarme it is assumed to be situated in the marls below the transgressive surface (Fig. 8) where black pebbles and charophytes occur. The lowstand deposits identified in the hemipelagic Montclus section comprise 900 kyr and, at Chapeau du Gendarme, are condensed to a few tens of centimetres (as seen in the Saleve section). At sequence boundary Be 5, at least 100 kyr are missing (corresponding to small-scale sequence 18; Fig. 8). The 25-85 m of non-decompacted section would thus represent the history of 1.7 Myr (Fig. 4), which gives an average preservation rate of 0.015 mm yr\(^{-1}\). This number is meaningful only when the long-term history of the Jura platform is discussed. When considering the (non-decompacted) sediment stack preserved and using the time span suggested by the cyclostratigraphical interpretation in Fig. 8, then the same 25-85 m would represent only 800 kyr, implying an average preservation rate of 0.033 mm yr\(^{-1}\). Still, this number is meaningless when discussing the rates of the sedimen-
The detailed sedimentological, sequence-stratigraphic and cyclostratigraphic analysis of Berriasian sections in the Swiss and French Jura reveals a complex but structured record of shallow-water carbonate facies. Their vertical evolution allows deepening-up and shallowing-up trends to be distinguished that define depositional sequences of different orders, the boundaries of which are furthermore enhanced by marly layers. The hierarchical stacking of these sequences and the time-frame given by bio- and sequence stratigraphic correlation with other, well-dated basins in Europe suggest that high-frequency, low-amplitude eustatic sea-level changes were a controlling factor and that they were related to insolation changes induced by the cyclical perturbations of the Earth’s orbit. The precession cycle (20 kyr) controlled the formation of the smallest units (elementary sequences), and the short eccentricity cycle (100 kyr) was responsible for small-scale sequences. These then group into medium-scale sequences related to the long eccentricity cycle (400 kyr). Although random processes such as lateral migration of sediment bodies or storm events were superimposed, the interpreted orbital cyclicity offers a time frame within which the durations of hiatuses and condensation intervals can be discussed, and sedimentation rates can be estimated.

As the encountered facies resemble those seen in modern shallow carbonate environments, average sediment accumulation rates can be assumed to be comparable. For the studied Berriasian example it is thus estimated that 44 kyr would have been sufficient to produce a sedimentary record of about 45 m, which corresponds today to about 26 m of compacted limestone and marls. However,
the time interval suggested by lateral correlation with numerically dated sequences is about 800 kyr. This means that the missing 756 kyr are lost in surfaces or condensed in thin sedimentary layers.

Short-term processes such as tidally controlled migration of ripples and dunes, scouring of tidal channels, or storm-induced erosion lead to hiatal surfaces where the record of a few hours to a few hundreds of years is lost. On the longer term, thousands to millions of years are not recorded or represented only by thin layers of sediment at sequence boundaries, at transgressive surfaces, and during maximum flooding. In the studied outcrops, reactivation surfaces have been identified within grainstones, whereas the boundaries of elementary and small-scale sequences in many cases display clasts that suggest reworking of previously cemented and now missing sediment. Iron staining and shifts in C and O isotopes point to subaerial exposure where again geological time has been lost. At the base of the studied sections, about 900 kyr are condensed in a few tens of centimetres of sediment, whereas at the top of the sections 100 to 300 kyr are not recorded. Such long-term gaps and condensations commonly are amalgamations or stacks of narrowly spaced shorter term hiatal surfaces. The absence of entire small-scale sequences and the lateral changes of thickness of the preserved sequences are attributed to synsedimentary tectonics that structured the Jura platform: lows favoured sediment accumulation while highs were prone to emergence followed by non-deposition or erosion.

Despite the many uncertainties related to numerical dating and the sequence- and cyclostratigraphic interpretations, this study allows the durations of hiatuses and condensations to be discussed in a narrow time frame on the order of a few tens of thousands of years. Within this time frame, facies evolution and hiatal surfaces of individual depositional sequences can be interpreted in detail, although additional uncertainties are introduced by the comparison with modern sediment accumulation rates. Nevertheless, it becomes clear that, when estimating sedimentation rates in the fossil record, the shortest possible time spans must be considered to minimize the effects of hiatuses and condensation. Although this is still far from the time resolution available in the Holocene and the Recent, a relatively realistic and dynamic picture of an ancient carbonate-dominated platform can be gained. The depositional record stays fragmentary, but the missing time can be well interpreted by comparison with processes seen today in modern carbonate systems.

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