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Carbonate Sedimentation Rates Today and in the Past: Holocene of Florida Bay, Bahamas, and Bermuda vs. Upper Jurassic and Lower Cretaceous of the Jura Mountains (Switzerland and France)

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Key words: Shallow-marine carbonates, Sedimentation rate, Sequence stratigraphy, Cyclostratigraphy, Holocene, Upper Jurassic, Lower Cretaceous, Florida Bay, Bahamas, Bermuda, Jura Mountains.

Abstract

Lagoonal to intertidal sediments from the Holocene in Florida Bay, on the Bahamas, and in Bermuda are compared to similar facies in the Kimmeridgian and Berriasian of the Swiss and French Jura Mountains. Dating by ^{14}C permits the estimation of sediment accumulation rates in the Holocene. In the ancient outcrops, the timing is given by cyclostratigraphic analysis. Elementary depositional sequences formed in tune with the 20-ka precession cycle, although much of this time may have been spent in non-deposition and/or erosion. After decompaction of the ancient sequences, their accumulation rates can be evaluated. It is suggested that the studied Holocene sediments accumulated over the past 6000 years with rates of 0.3 to 3 mm/a, whereas the Kimmeridgian and Berriasian facies show somewhat lower rates of 0.07 to 0.6 mm/a. This difference may be due to methodological errors, but also to variable carbonate production. In shallow carbonate systems, much of the sediment produced may be redistributed over the platform or exported. Furthermore, basin morphology and currents can strongly influence facies and thickness of the accumulated sediment. Finally, the accumulated sediment may suffer further erosion before it is preserved in the sedimentary record. Consequently, when estimating sedimentation rates in ancient sequences, it is important to do this with the highest time resolution possible, and only after having decompacted the sediment and evaluated the time lost in hiatuses or condensed intervals.

1. INTRODUCTION

Knowledge about rates of sedimentation is of great importance for the understanding of sedimentary systems. Sedimentation rates reflect the availability of sediment and the conditions in the area of accumulation. Sedimentation rates also are important input parameters for computer modelling.

Sediment accumulation is controlled by supply (i.e. material transported into the system and/or produced in-situ) and accommodation (i.e. the space available to store this material – SCHLAGER, 1993). Sediment **supply rates** may be highly variable, from rapid coral-reef growth to pelagic fall-out controlled by seasonal plankton blooms, and to episodic deposition through storms or turbidity currents. Rates of accommodation change are the combined rates of sea-level change and change in subsidence. In shallow water, supply may be higher than accommodation, and sediment is redistributed into other environments. It is therefore useful to speak of **accumulation rate** in a given position in the sedimentary basin. Once the sediment is accumulated, it may be eroded and redistributed through changes in accommodation and/or current activity. The final stratigraphic record commonly represents only part of the sedimentation history, and it might therefore be adequate to distinguish also **preservation rate**. Several authors (e.g., BOSSCHER & SCHLAGER, 1993; PLOTNICK, 1986; SADLER, 1981, 1994; SCHLAGER, 1999) have argued that "accumulation" rates decrease with increasing time spans over which they are calculated. This relationship probably reflects the fact that more hiatuses are included with increasing time intervals (i.e. a diminished preservation potential) rather than changes in sediment availability.

In this paper, we concentrate on sedimentation rates on shallow carbonate platforms where sediment is to a large extent produced by organisms. Consequently, ecological control is important. Siliciclastics and nutrients are introduced by continental run-off, which in turn is controlled by climate. By first evaluating sedimentation rates in Holocene systems where the controlling parameters are relatively well known, we then estimate rates in ancient sedimentary environments. We have chosen Upper Jurassic and Lower Cretaceous sequences that contain facies comparable to those studied in the Holocene and where high-resolution sequence stratigraphy and cyclostratigraphy furnish the adequate time frame.

2. HOLOCENE SEDIMENTATION RATES

A large amount of literature is available where Holocene sedimentation rates in shallow-water carbonate systems are indicated (e.g., CLOUD, 1962; SHINN et

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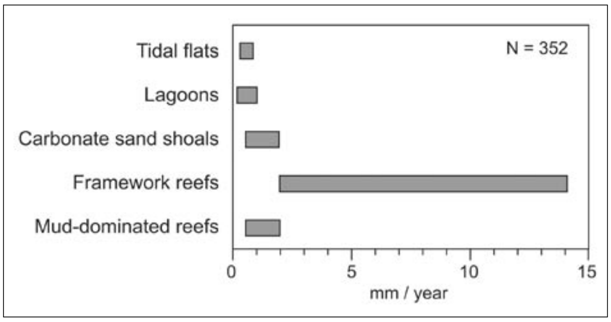

Fig. 1 Compilation of 352 sedimentation rates according to the literature cited in the text.

al., 1965; STOCKMAN et al., 1967; KUKAL, 1971, 1990; SMITH, 1971; CHAVE et al., 1972; THOMSON & TUREKIAN, 1973; ENOS, 1974, 1991; MOORE, 1972; SARNTHEIN, 1973; TURMEL & SWANSON, 1976; GEBELEIN, 1977; HALLOCK, 1981; BOSENCE et al., 1985; WANLESS & TAGETT, 1989; KENNEDY & WOODROFFE, 2000; HEAP et al., 2001; MACKINNON & JONES, 2001; CALHOUN et al., 2002; YANG et al., 2002; GISCHLER, 2003). In Fig. 1, 352 data points assembled from the references cited above have been grouped to represent different depositional environments. Coral framework reefs are the fastest sediment producers, whereas sediment on tidal flats and in lagoons accumulates relatively slowly. Ooid, peloid, or bioclastic shoals accumulate within a relatively small range of 0.5–2 mm/a.

In order to also evaluate lateral changes in facies and accumulation rates, shallow cores have been taken in Florida Bay, on the Bahamas, and in Bermuda. We have chosen sites that are mostly protected from high energy and constant reworking, and where relatively continuous sediment accumulation can be expected. A plastic tube was pushed or hammered into the soft sediment, sealed, and then pulled back. The sediment core was extracted with a pushing device, and the total compaction induced by these processes was measured. It is assumed that this compaction partly reflects the mechanical compaction the sediment would undergo through burial. In fact, the calculated compaction corresponds well to the values obtained experimentally by SHINN & ROBBIN (1983) from similar facies. The cores were described and sampled. Samples were impregnated with epoxy, and thin sections were made. Mangrove peat, bivalve shells, or corals were taken for ^{14}C analyses. Sample preparation and dating was performed by the radiocarbon laboratory of the Institute of Physics at the University of Berne, Switzerland. The results are listed in Table 1. The ages are conventional (calculated by using the Libby half-life value) and expressed in years before present (1950).

Locality, sample number (lab number)	*Material dated*	*C^{14} age (a BP)*	*Depth below sea floor (not compacted) (cm)*	*Major facies in core*	*Average accumulation rate (mm/a)*
Crane Key (Florida)					
CK-1B (B-5407)	peat	1040±60	90	peat, mud, some sand	0.9
CK-2B (B-5408)	peat	1070±40	85	peat, mud, some sand	0.8
CK-3B (B-5409)	peat	1360±60	50	peat, mud, sand	0.4 (erosion)
Cotton Key (Florida)					
Flo-4 (B-7792)	bivalve shells	1040±100	115	mud, shells	1.1
Flo-6 (B-7793)	corals	680±110	90	mud, roots	1.3
Pigeon Key (Florida)					
Flo-19 (B-7807)	roots	2320±90	85	peat, mud	3.0 (up to Flo-22)
Flo-22 (B-7794)	peat	2140±30	30	peat, mud	0.1 (erosion)
Little Crawl Key (Florida)					
Flo-23a (B-7795)	corals	1210±30	75	sand, coral gravel	0.6 (winnowing)
Flo-23c (B-7797)	corals	1210±40			
Three Creeks (Bahamas)					
Bah-6 (B-7801)	corals	3720±70	140	mud, roots	0.4
Lee Stocking (Bahamas)					
LSC-1 (B-5411)	peat	3440±90	120	peat, sand, mud	0.3
LSC-2 (B-5412)	peat	2530±120	85	peat, sand, mud	0.3
The Lagoon (Bermuda)					
Ber-9 (B-7427)	shell debris	2490±30	105	mud, some sand	0.4
Ber-19 (B-7429)	shell debris	5630±40	65	sand, mud	0.1 (lag deposit)
Tucker's Town (Bermuda)					
Ber-27 (B-7433)	shell debris	2840±30	130	sand	0.5

Table 1 Dated samples and estimated sediment accumulation rates for the Holocene cores. For interpretation see text.

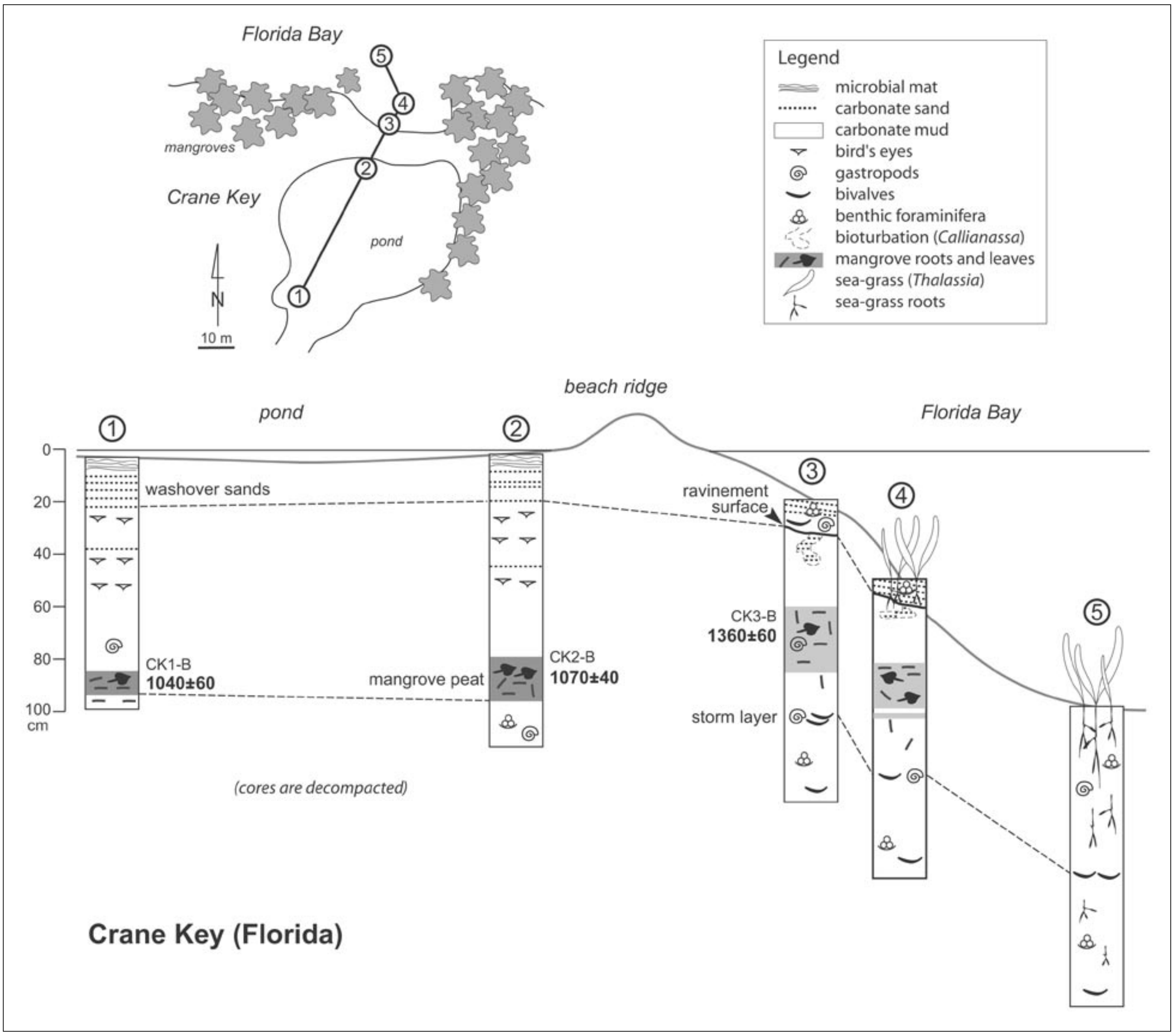

Fig. 2 Shallow sediment cores on Crane Key, Florida Bay (to the north of Windley Key).

2.1. Florida Bay

The Holocene lagoonal sediments and mudmounds in Florida Bay are well described by, e.g., ENOS & PERKINS (1979), WANLESS & TAGETT (1989), and BOSENCE (1989a, b, 1995). Sedimentation started when post-glacial sea-level rise induced flooding of the subaerially exposed Pleistocene limestones and created accommodation space. The Florida Keys protect the Bay from high energy, and mostly muddy, biodetrital carbonates accumulate. Mud banks aggrade and prograde. Islands form, which are colonised by mangroves and locally contain freshwater lakes. Windward (i.e. northern to eastern) margins of the islands are commonly eroded, whereas on the leeward sides mud accumulates in elongated banks (ENOS & PERKINS, 1979).

On Crane Key, five cores have been taken along a transect from a pond on the island, over a beach ridge, to the shallow lagoon (Fig. 2). Below the peat in cores 1 and 2, a lagoonal facies of grey carbonate mud with peloids, benthic foraminifera, bivalves, and gastropods is encountered. This same facies dominates in core 5, where roots of sea-grass are also common. Mangrove peat in cores 3 and 4 implies that the shoreline was positioned farther to the north than today. It is interesting to note that the mangrove peat below the pond (in cores 1 and 2) is younger but topographically lower than the peat in the shoreface core (3). This suggests that a shallow mangrove pond existed behind a beach barrier already 1000 years ago. Above the peat in cores 1 to 4, yellowish mud with very little fauna predominates, indicating a restricted environment. Bird's eye structures occur in cores 1 and 2 and indicate the intertidal zone. Ravinement surfaces appear in cores 3 and 4, which are overlain by sand composed of peloids, bivalves, gastropods, and foraminifera. Layers of this sand are found in the pond behind the beach ridge and imply washover processes. Microbial mats finally seal the cores in the pond.

On Cotton Key, three cores represent a mud bank, and one core a small bay suffering erosion (Fig. 3). The dominant facies in all cores is carbonate mud with

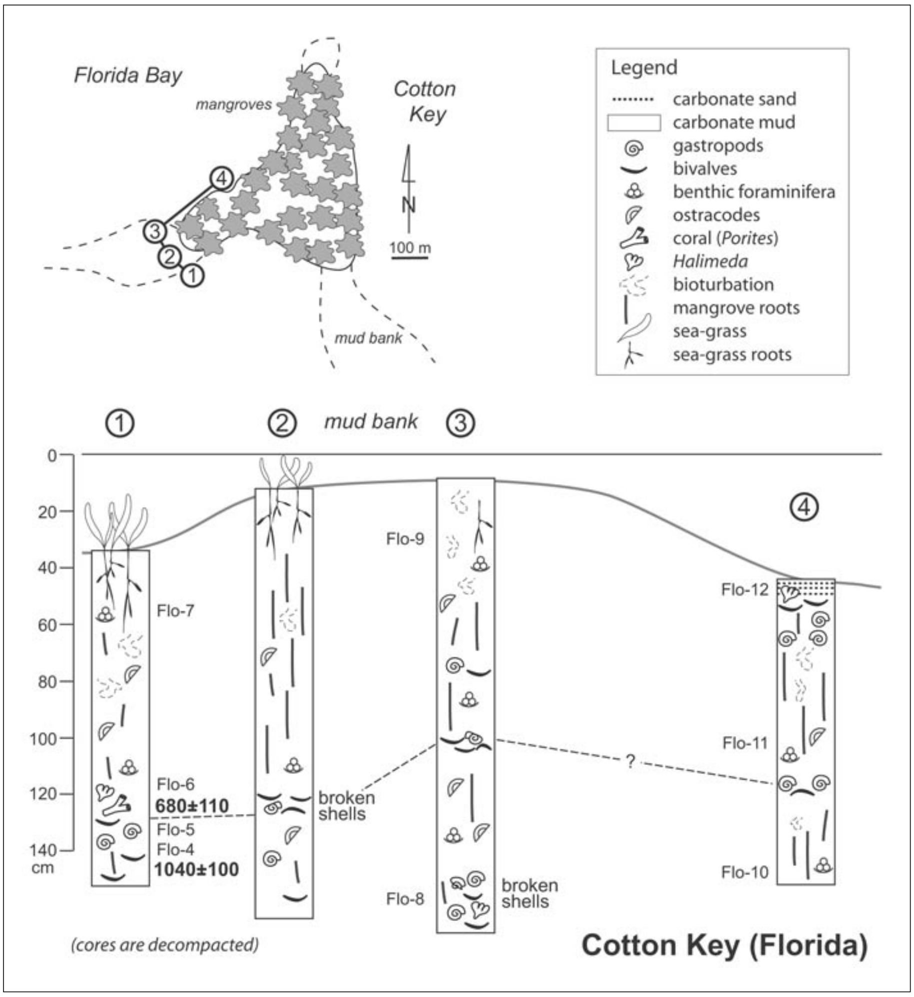

Fig. 3 Shallow sediment cores on Cotton Key, Florida Bay (to the northwest of Windley Key).

peloids, benthic foraminifera, bivalves, gastropods, ostracodes, and fragments of the green alga *Halimeda*. Layers with concentrations of broken shells indicate periodic reworking by high energy. A *Porites* coral fragment points to storm transport through a channel from the southern side of the Florida Keys, where coral carpets occur (BOSENCE et al., 1985). On the top of core 4, the sediment is winnowed to produce peloidal–bioclastic carbonate sand.

Pigeon Key has furnished two cores (Fig. 4). One is situated on the erosive eastern side of the island and displays two phases of mangrove growth. Carbonate sand is winnowed on the sediment surface and preserved in crab burrows. The second core represents the mud bank. Peloids, bivalves, gastropods, benthic foraminifera, ostracodes, and organic fragments float in a muddy matrix.

Little Crawl Key is situated on the southern side of the Florida Keys (Fig. 5). Energy generally is higher than in Florida Bay, and a sandy beach has developed. Coral fragments (*Porites*) at the base of the cores indicate fully marine conditions. Winnowed peloidal sands with gastropods, bivalves, *Halimeda*, red algae, and benthic foraminifera dominate the facies. Muddy sediment with bioturbation occurs in the middle part of the cores.

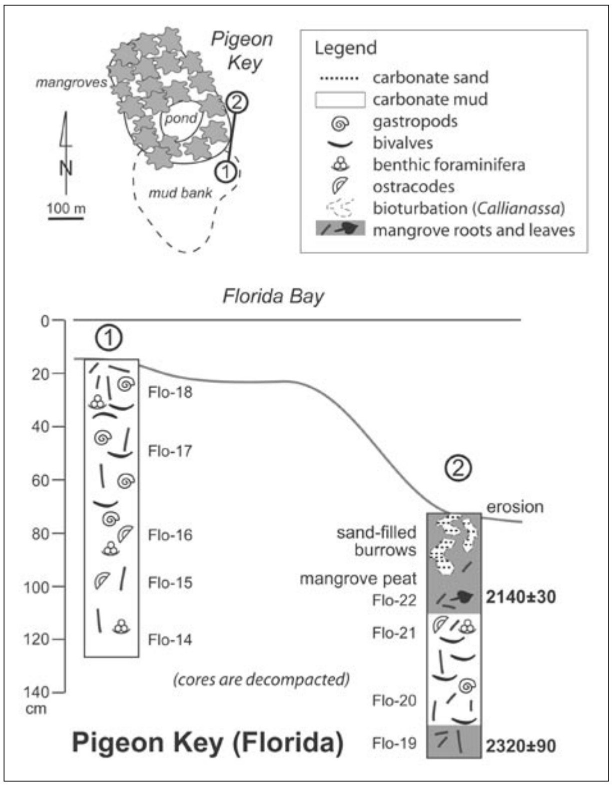

Fig. 4 Shallow sediment cores on Pigeon Key, Florida Bay (north of Tavernier on Key Largo).

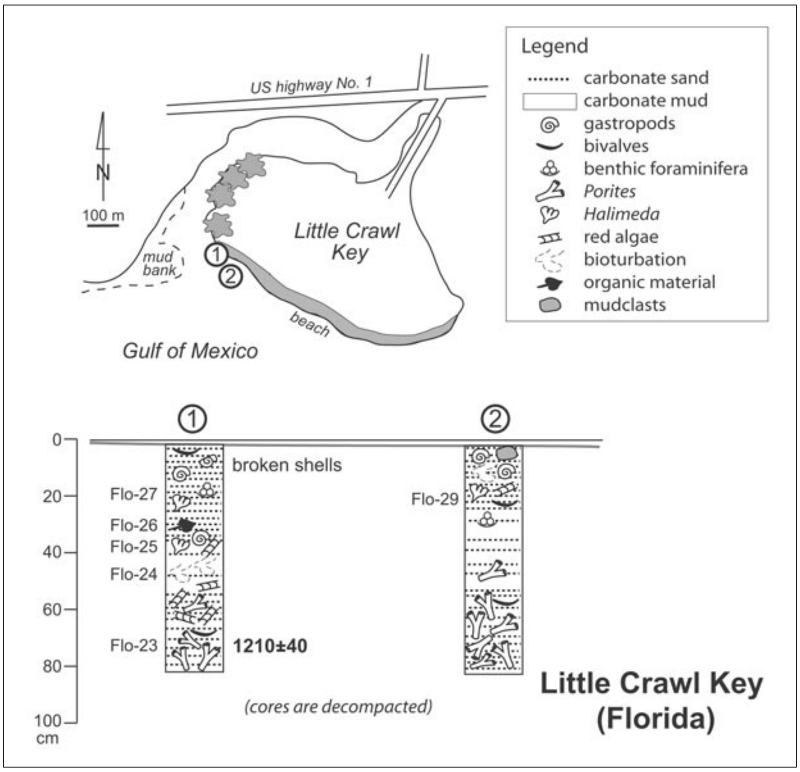

Fig. 5 Shallow sediment cores on Little Crawl Key (south of US highway No. 1, west of Crawl Key).

2.2. Bahamas

We have taken cores in two low-energy settings from the Bahamas: on Andros Island, and on Lee Stocking Island. The Holocene sedimentation of the Andros tidal flats has been described in great detail by HARDIE (1977).

In the Three Creeks area on Andros Island, a core taken on the levee of a tidal channel reveals mudclasts in its lower part (Fig. 6). This suggests lateral migration of the levee facies over the channel floor where such clasts had accumulated. The core taken in the pond shows a relatively homogeneous facies of carbonate mud with peloids, benthic foraminifera, bivalves, and gastropods. The gastropods (Cerithids) are concentrated locally but are in-situ.

On Lee Stocking Island (Exuma Cays), an isolated pond has been cored (Fig. 7). The mangrove peat in the lower part of core 1 accumulated over a period of about 1000 years. It is followed by sandy facies containing peloids, bivalves, Cerithid gastropods, benthic foraminifera, and ostracodes. The upper part of the core is muddy but still contains Cerithids. In core 2, which now is closer to the border of the pond, marine facies including echinoids directly overlie the Pleistocene substrate. In its upper muddy part, foraminifera still testify to a marine influence. It is concluded that this pond has evolved from a bay, and that it was closed by a beach barrier only very recently.

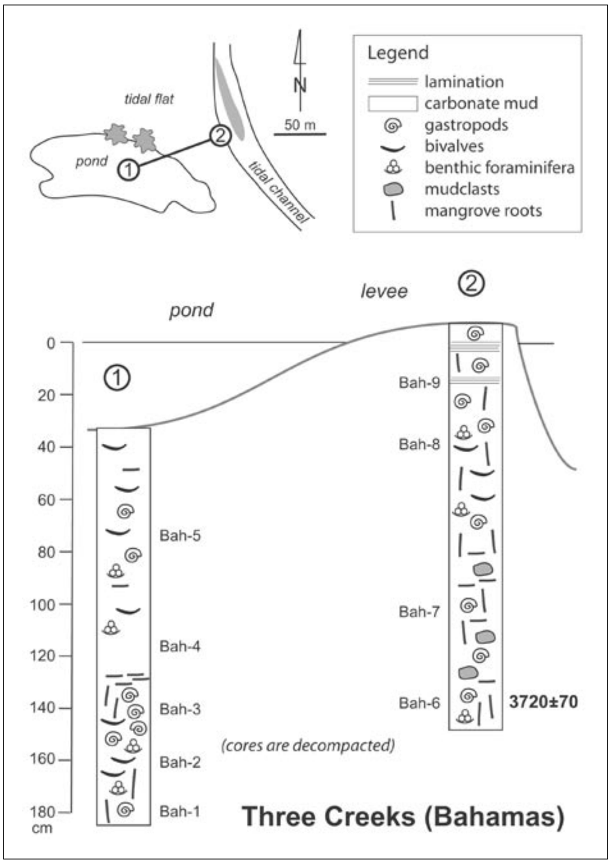

Fig. 6 Shallow sediment cores on the southernmost tidal channel in the Three Creeks area, northwestern Andros Island, Bahamas.

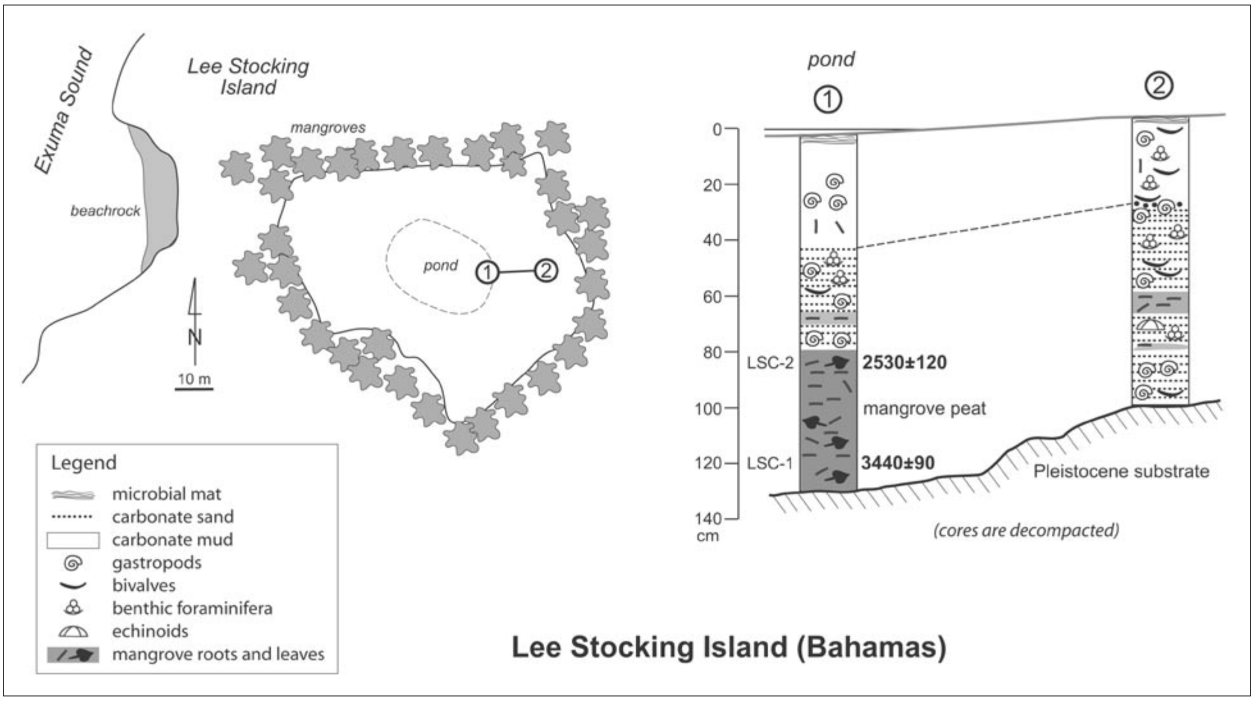

Fig. 7 Shallow sediment cores in a pond on the northwestern tip of Lee Stocking Island, Exumas, Bahamas.

2.3. Bermuda

Two sites have been studied on Bermuda: The Lagoon on Ireland Island South and Tucker's Town Bay on the southern side of Castle Harbour.

The Lagoon is connected to the ocean by two narrow inlets on either side (Fig. 8). Five cores have been taken, which show periodic high-energy influence in a generally low-energy system. Facies are marine, including peloids, bivalves, gastropods, benthic foraminifera, and abundant *Halimeda* plates. Some bioclasts are blackened, indicating locally anoxic conditions (STRASSER, 1984). Roots occur in all cores, and pieces of wood are locally concentrated in layers. Lithoclasts indicate some reworking. The activity of the washover bar (cores 2 to 4) might have been somewhat influenced by the construction of the road along Great Sound. The relatively high age of the material dated at the base of core 5 in Great Sound indicates storm reworking close to the Pleistocene substrate.

In Tucker's Town Bay, peloidal–bioclastic carbonate sands including the red foraminifera *Homotrema rubrum* fill the southern branch of the bay (Fig. 9). The cores start with fine sands, which become coarser towards the top.

2.4. Holocene accumulation rates

Because only few samples are suitable for ^{14}C dating, accumulation rates can only be calculated as averages over a few thousand years. Additional errors come from facies-dependent differential compaction (e.g., carbonate mud versus peat), which has not been corrected for. However, if the facies in a core is more or less homogeneous and no erosion or reworking occurred, it can be assumed that the values shown in Table 1 are representative.

Accumulation rates for the studied low-energy, lagoonal, and peat sediments vary between 0.3 and 3.0 mm/a. These values are comparable to the ones compiled in Fig. 1. The sandy facies in Tucker's Town Bay accumulated with an average of 0.5 mm/a. Deposition of muddy sediment was precluded by tidal currents. On Little Crawl Key, coral rubble at the base of the cores and beach sands at the top imply some winnowing. Nevertheless, because the sandy sediment in Tucker's Town Bay and on Little Crawl Key filled in the available accommodation space and no erosion surfaces are visible, the estimated sedimentation rates can be considered as accumulation rates. Low values appear where there is evidence for erosion (on Crane and Pigeon Keys) or for a lag deposit (in Great Sound, Bermuda). These rates therefore do not express sediment production and accumulation but rather the potential for preservation.

3. HIGH-RESOLUTION SEQUENCE STRATIGRAPHY AND CYCLOSTRATIGRAPHY

If sediment accumulation is controlled by sea-level fluctuations, the concepts of sequence stratigraphy can be applied to describe and interpret the resulting sedimentary sequences. Although sequence stratigraphy classically deals with large sequences lasting a few hundred–thousand to a few million years ("third-

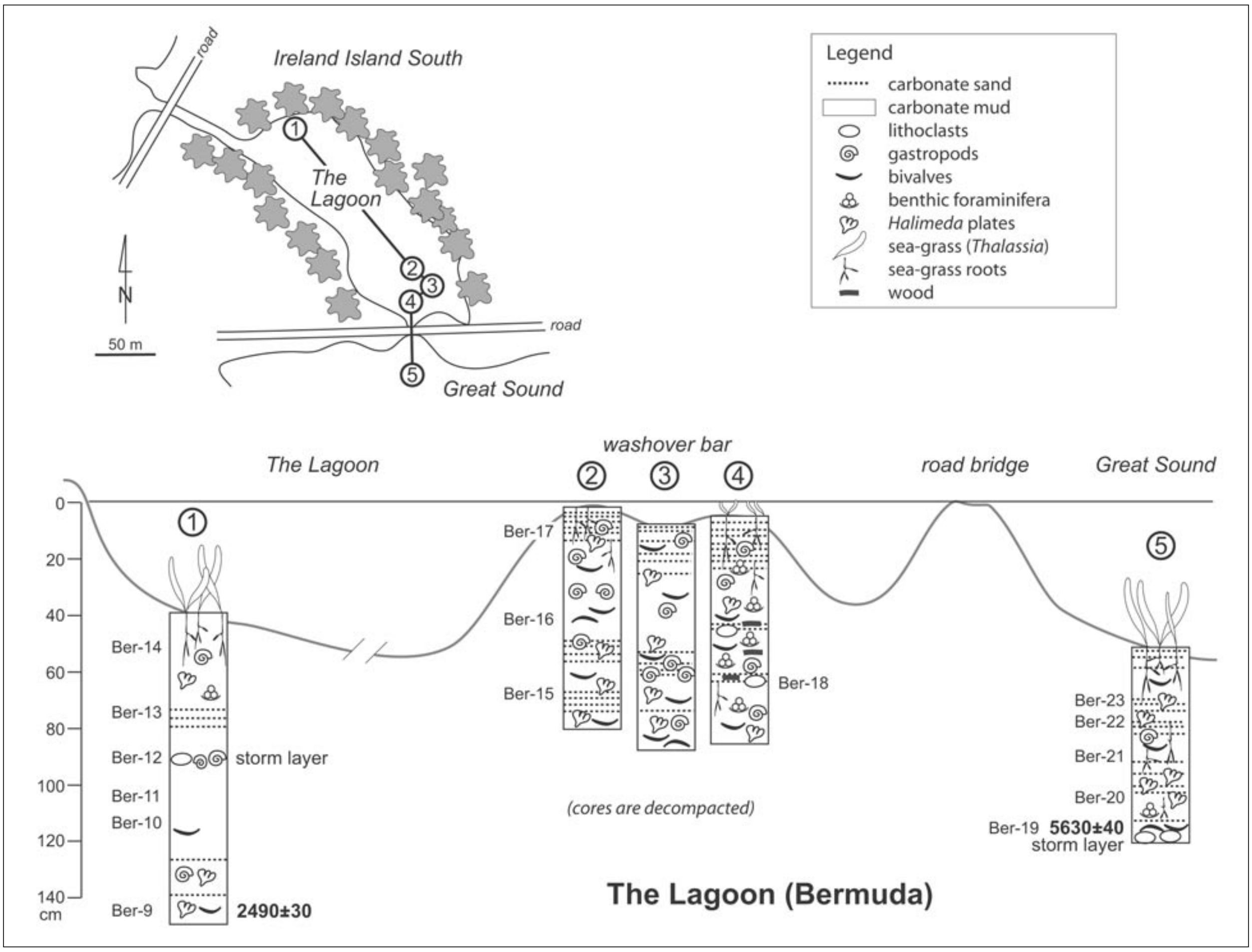

Fig. 8 Shallow sediment cores in The Lagoon (Ireland Island South, western Bermuda).

order" sequences of VAIL, 1987; VAIL et al., 1991), its terminology can also be applied to deposits forming in much shorter time intervals (e.g., MITCHUM & VAN WAGONER, 1991; POSAMENTIER et al., 1992; STRASSER et al., 1999).

According to NEUMANN (1971), DIGERFELDT & HENDRY (1987), and BOARDMAN et al. (1989), sea level in the Caribbean and Bermuda rose rapidly from its last glacial lowstand until about 5000 years BP and then slowed down to reach its present position. This curve represents the rising limb of a sea-level cycle controlled by insolation changes coupled to the orbital precession cycle of 20 ka (BERGER et al., 1989). Consequently, the Holocene sediments on shallow carbonate platforms are the record of only part of a sea-level cycle (Fig. 10).

In the Pleistocene, when sea level was mainly controlled by slowly waxing and rapidly waning ice caps, the sea-level curve especially of the 100-ka eccentricity cycles was highly asymmetric (e.g., SHACKLETON, 1987). In past greenhouse worlds, ice in high latitudes probably was present (FRAKES et al., 1992; EYLES, 1993; VALDES et al., 1995), but ice-volumes were not sufficient to induce important glacio-eustatic fluctuations. However, volume changes of alpine glaciers could make a small contribution (FAIRBRIDGE, 1976;

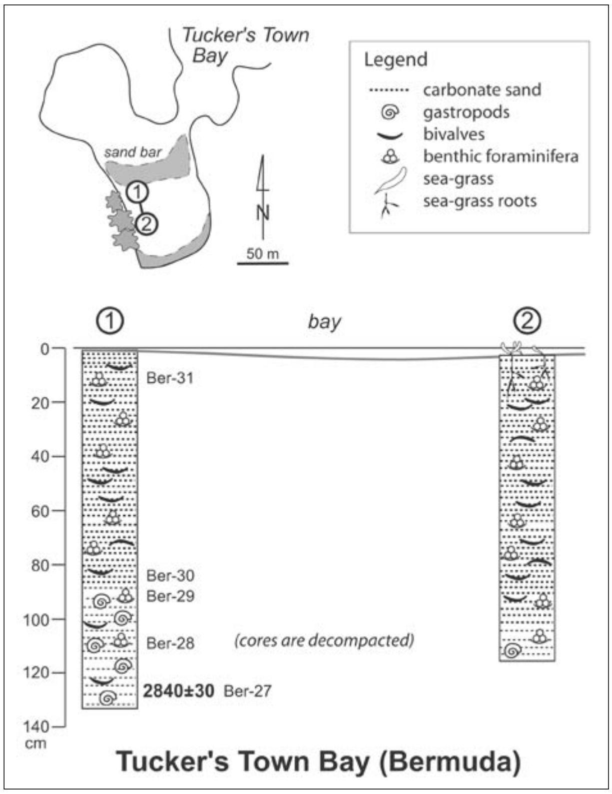

Fig. 9 Shallow sediment cores in Tucker's Town Bay, eastern Bermuda.

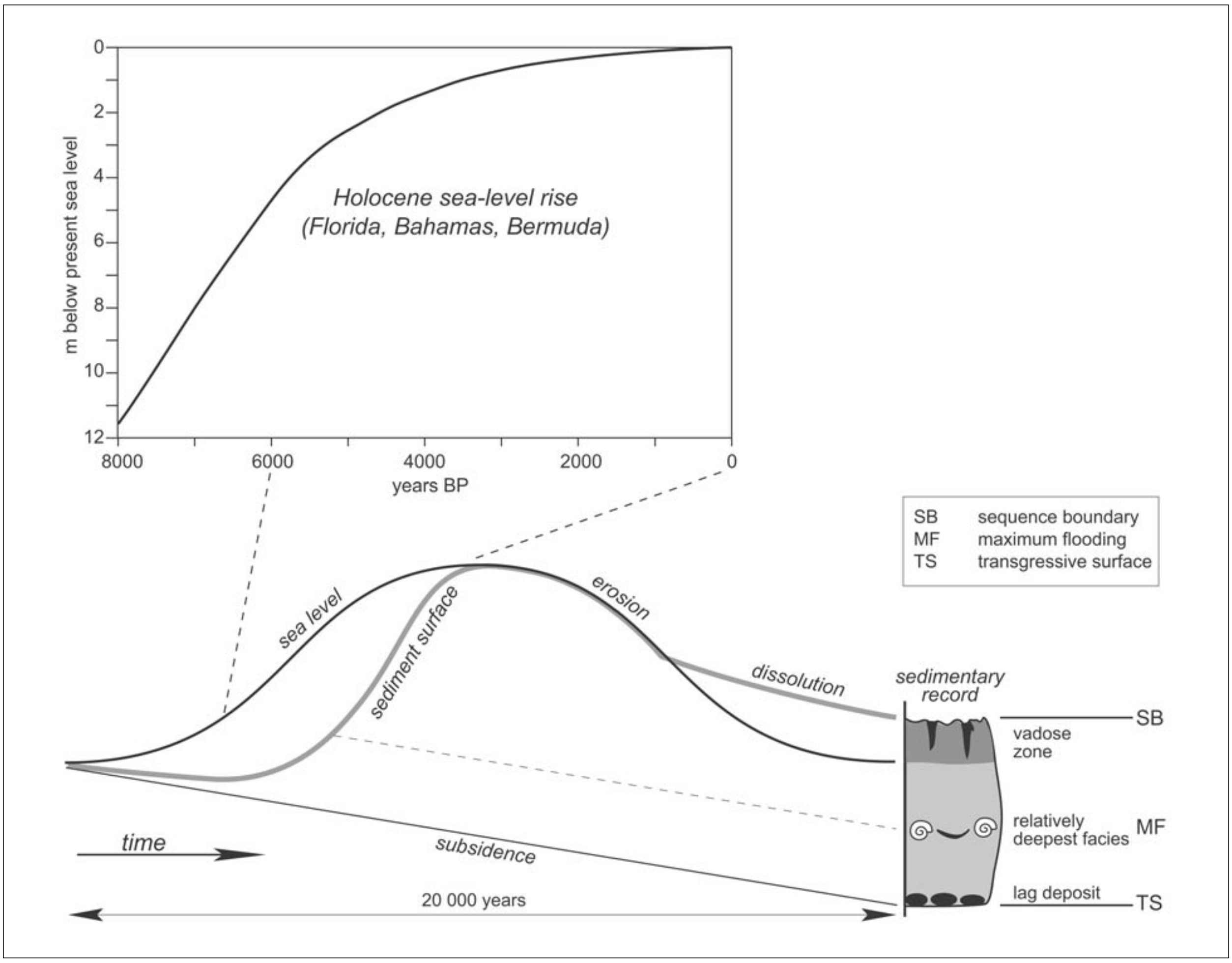

Fig. 10 Trend of Holocene sea-level rise and comparison with a hypothetical, symmetrical sea-level cycle related to the 20-ka orbital precession cycle. For discussion see text.

VALDES et al., 1995). Sea-level changes were also created by thermal expansion and retraction of the uppermost layer of ocean water (GORNITZ et al., 1982), by thermally-induced volume changes in deep-water circulation (SCHULZ & SCHÄFER-NETH, 1998), and/or by water retention and release in lakes and aquifers (JACOBS & SAHAGIAN, 1993). Consequently, sea-level cycles during greenhouse conditions were of low amplitude and probably relatively symmetrical (READ et al., 1995).

On shallow carbonate platforms, initial flooding of a subaerially exposed surface results in a lag deposit. Sediment production starts up once the carbonate-producing organisms have colonised the newly available space. If the ecological conditions are suitable, carbonate production will soon outpace sea-level rise (NEUMANN & MACINTYRE, 1985) and fill in the available space. Slowing-down of sea-level rise will further accelerate this process, and sea-level drop will lead to erosion and reworking of the previously deposited sediment. If a fresh-water lens develops, carbonate cementation sets in within a few hundred years (e.g., HALLEY & HARRIS, 1979) and stabilises the sediment. Karstification may further lower the sediment surface (Fig. 10). The resulting sedimentary record thus shows first a deepening, then a shallowing trend of facies evolution. In the sequence-stratigraphic terminology, the initial flooding corresponds to the transgressive surface, the relatively deepest facies to the maximum flooding, and the erosion surface to the sequence boundary.

The sediment cores illustrated in Figs. 2 to 9 would thus correspond to the late highstand deposits, covering the last few thousand years of sedimentation history. An exception is core 5 of The Lagoon in Bermuda (Fig. 8), where the shell debris dated at about 5600 years BP may represent a storm-influenced transgressive lag. Locally, the sediment has filled the available accommodation space and even built up low islands, while in other places the sediment surface still is subtidal. Current- and wave-induced erosion is active in the case of the mounds in Florida Bay, but cementation and karstification due to sea-level fall has not yet occurred. Global warming is even causing sea level to rise again (IPCC, 2001), thus counteracting the fall that would be expected according to the orbitally induced reduction in insolation (BERGER, 1978).

In the ancient sedimentary record, depositional sequences commonly are hierarchically stacked (e.g., GOLDHAMMER et al., 1993; MONTAÑEZ & OSLEGER, 1993; D'ARGENIO et al., 1997). STRASSER et al. (1999) proposed a descriptive classification of elementary, small-scale, medium-scale, and large-scale sequences. An elementary sequence is the smallest detectable unit where facies changes indicate one cycle of environmental change (including accommodation change). For the estimation of sedimentation rates, it is mandatory to analyse the smallest units that can be distinguished, i.e. the elementary sequences, which also correspond to the shortest time span. By detailed analysis of facies and surfaces, short-term changes in accumulation rate and hiatuses may be detected, and a more realistic picture of the sedimentary history is obtained than by averaging over thick sequences and long time spans.

The error margins for dating ancient sedimentary rocks are large and can reach values on the million-year scale (BERGGREN et al., 1995). However, a relatively high time resolution can potentially be reached by cyclostratigraphy. The periodicities of the orbital cycles (Milankovitch cycles) are known (e.g., SCHWARZACHER, 1993). Although the motions within the solar system are chaotic and the predictability of the orbits of the inner planets (including Earth) is lost within a few tens of millions of years (LASKAR, 1989), the periodicities can also be estimated for older geological times (BERGER et al., 1989). If it can be shown that the sedimentary record was controlled by orbitally induced environmental changes (including sea-level changes), then a relatively high time resolution can be reached (potentially 20 ka). Through the sequence-stratigraphic interpretation, this time interval can be further subdivided (Fig. 10). The exact durations of the transgressive and highstand intervals, the transgressive lag time, and the time spent in erosion and non-deposition are, of course, speculative.

4. ANCIENT SEDIMENTATION RATES

In order to compare Holocene and ancient sedimentation rates, depositional sequences with facies similar to those described above have been chosen in Upper Jurassic and Lower Cretaceous sections in the Swiss and French Jura Mountains. The Jura realm at those times was a large, structurally complex, subtropical carbonate platform at the northern margin of the Tethys ocean (ZIEGLER, 1988; DERCOURT et al., 2000).

4.1. Facies and sequences

The Lower Berriasian (Purbeckian) is characterised by shallow-lagoonal, intertidal, and supratidal carbonate facies (CAROZZI, 1948; HÄFELI, 1966; STRASSER, 1988). The facies evolution through time allows the identification of depositional sequences, the formation of which was significantly influenced by low-amplitude, high-frequency sea-level changes. The biostratigraphic and chronostratigraphic framework, as well as the stacking pattern of these sequences, suggest that the sea-level fluctuations were controlled by orbitally induced insolation changes. Detailed descriptions and interpretations of the Purbeckian sections are given in STRASSER (1988) and STRASSER & HILLGÄRTNER (1998). The high-resolution sequence stratigraphy and cyclostratigraphy of the Kimmeridgian in the Swiss Jura have been studied in detail by COLOMBIÉ (2002). As in the Purbeckian, facies indicate shallow-lagoonal to peritidal depositional environments.

In the sequences shown in Figs. 11 and 12, lagoonal packstones to wackestone with normal-marine fauna dominate. Lithoclasts and black pebbles occur at the base of some sequences, indicating that the top of the previous sequence had been subaerially exposed and cemented before the transgression reworked these elements. The tops of the sequences commonly display bird's eyes and dolomitization implying intertidal to supratidal conditions. Erosion surfaces occur at the top of some sequences. Clay seams or thin marly layers separate the beds.

A comparison of Holocene and Kimmeridgian facies is shown in Fig. 13. The peloidal–bioclastic muds of the Holocene cores compare well with the lagoonal wackestones of the Kimmeridgian and Berriasian. In the ancient rocks, of course, diagenesis is more advanced and has led to the local development of dolomite crystals and to the filling of pore space with calcite cement.

4.2. Decompaction

Before sedimentation rates in the ancient rocks can be estimated and compared to their Holocene counterparts, the depositional sequences have to be decompacted. Mechanical reorganisation of grains and dewatering in carbonate mud leads to a porosity loss of 10 to 30% after the first 100 m of burial (MOORE, 1989). The experiments of SHINN & ROBBIN (1983) yielded values of 20 to 70% of volume loss through mostly mechanical and dewatering compaction. These values are lower if carbonate cementation sets in very early (e.g., HALLEY & HARRIS, 1979). With deeper burial, chemical compaction becomes important. Pressure solution at grain contacts testifies to some dissolution processes in the studied sediments. Their burial depth is not known but probably never exceeded 2 km (TRÜMPY, 1980). GOLDHAMMER (1997) proposes a compaction of slightly over 50% for 1 m of carbonate mud buried at 1000 m, and of about 15% for 1 m of carbonate sand at the same burial depth. According to ENOS (1991), muddy terrigenous and muddy carbonate sediments do not have significantly different compaction curves. However, pressure solution along clay seems may of course enhance chemical compaction in carbonates (e.g., BATHURST, 1987). Based on these published values, the following decompaction factors have been

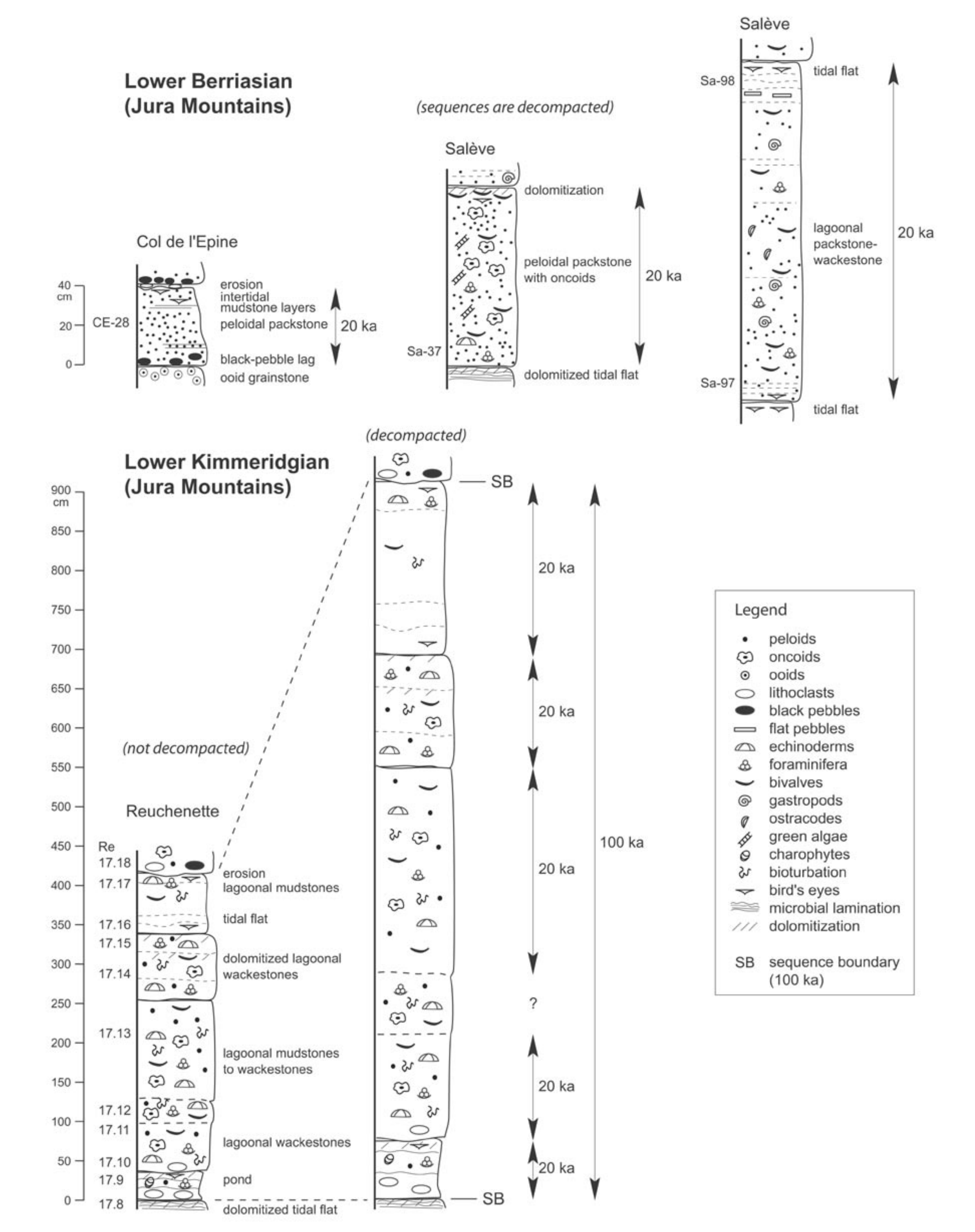

Fig. 11 Selected 20-ka sequences in Lower Berriasian and Lower Kimmeridgian sections of the Swiss and French Jura. For discussion see text.

chosen for the present study: 1.2 for grainstones and 2.5 for mudstones. For packstones and wackestones, the intermediate factors 1.5 and 2 are assumed, for marls the factor 3. These values are, of course, only rough estimates but nevertheless allow for an approximation of the original sediment thickness.

4.3. Estimation of time and accumulation rates

In Fig. 11, three individual elementary sequences from the Lower Berriasian are presented. They are interpreted to have formed in tune with Earth's precession cycle of 20 ka (STRASSER, 1988; STRASSER &

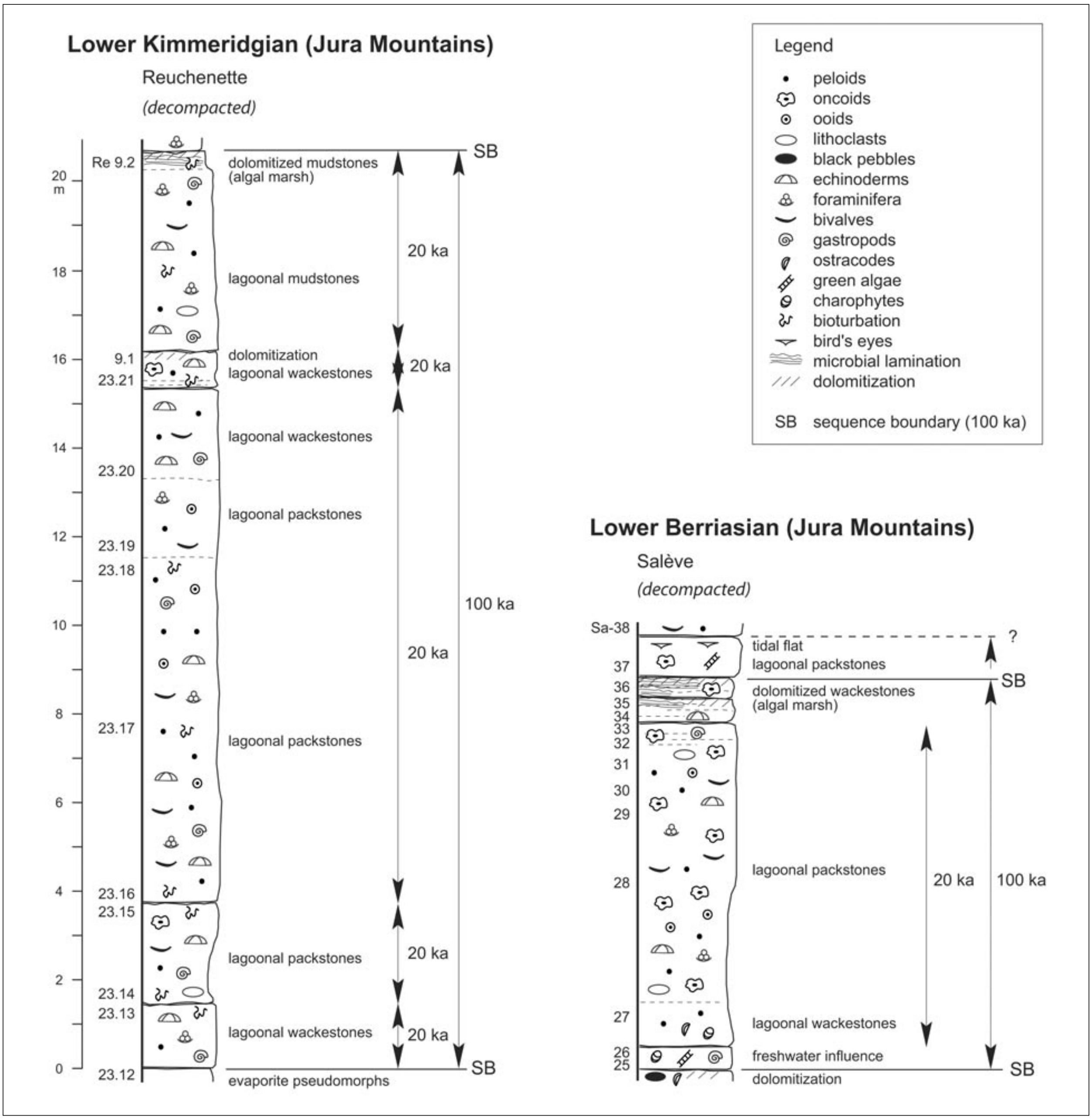

Fig. 12 The two thickest 20-ka sequences encountered in the studied Lower Kimmeridgian and Lower Berriasian sections, respectively. In the Salève section, only four 20-ka sequences can be interpreted within the 100-ka sequence. There may have been condensation at either one of the 100-ka sequence boundaries, or the 100-ka cycle terminates with the tidal flat below sample Sa-38.

HILLGÄRTNER, 1998). The sequence from the Col de l'Epine outcrop in the French Jura presents a lag deposit at its base and erosion features at its top. The tops of the two Salève sequences display bird's eyes, and one is dolomitized. In accordance with the model of Fig. 10 it is therefore suggested that only part of the 20 ka was actually spent in sediment accumulation. An estimate of 6000 years is used for the calculation of sedimentation rates (Table 2).

The part of the Kimmeridgian Reuchenette section shown in Fig. 11 illustrates the stacking of 20-ka sequences into a 100-ka sequence, the latter reflecting the first eccentricity cycle of the Earth's orbit (COLOMBIÉ, 2002). The lowermost 20-ka sequence is thin and shows reworking at its base and intertidal conditions at its top. The following sequences are thicker and display only subtidal facies. Marly seams separating the beds are probably related to sea-level falls that mobilized clays in the hinterland but did not cause emersion at the site where the observed sediments were deposited. At the level of sample Re-17.12, high accommodation precluded distinct features that mark a boundary between sequences. From lateral correlation it is implied that two 20-ka sequences compose the interval from sample Re-17.10 to 17.13, but it is not clear which one of the two joints (above and below Re-17.12) corresponds to the boundary. In sample Re-17.15, subtidal

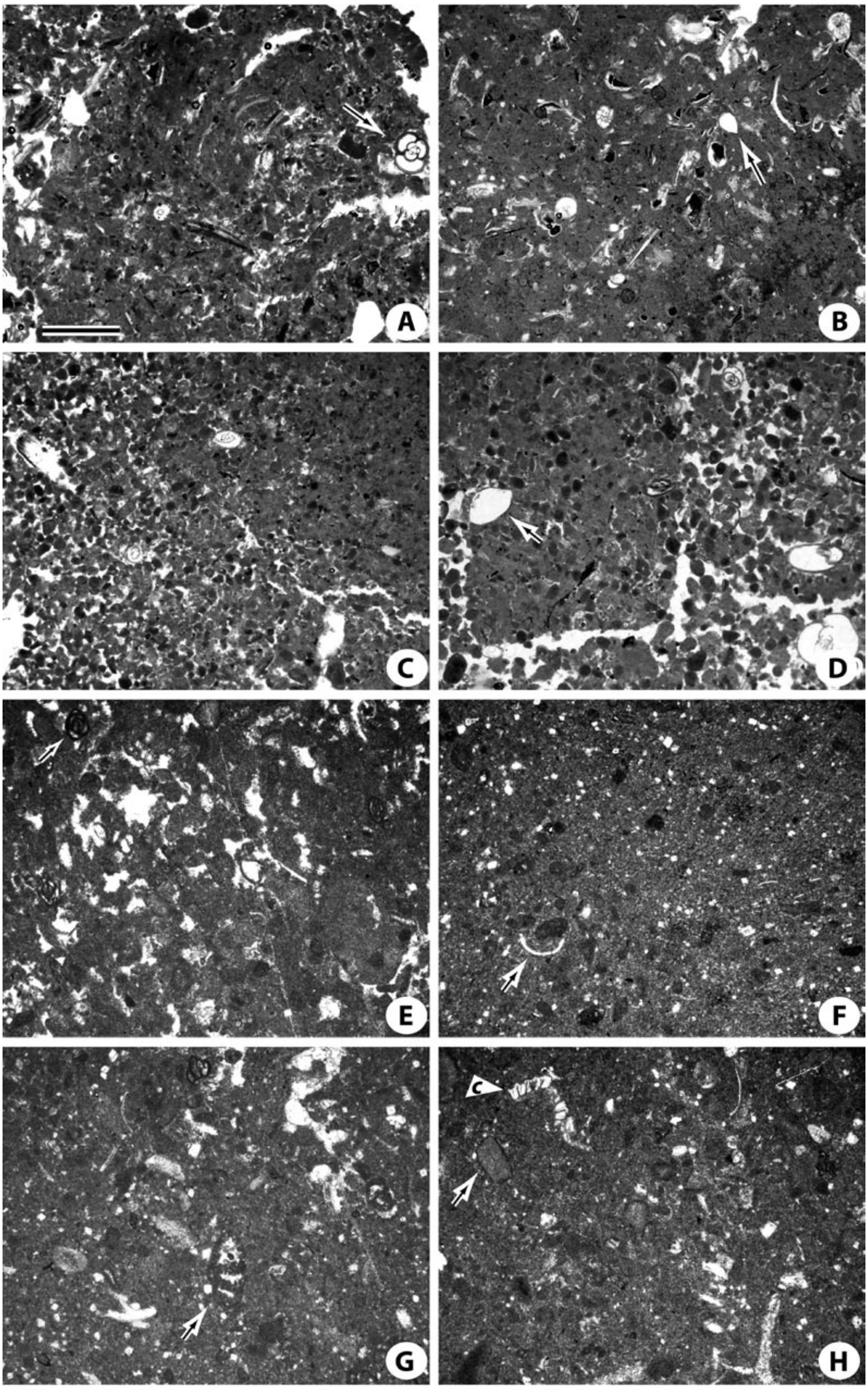

Fig. 13 Comparison of Holocene and Lower Kimmeridgian facies. The bar in A is 0.5 mm long and valid for all photomicrographs.

A–D Holocene material (fractures are due to sample preparation)**:** **A** – Peloidal wackestone. The biotic components are benthic foraminifera (arrow) and bivalves (sample Flo-5). **B** – Wackestone. Ostracodes (arrow), benthic foraminifera, shell fragments, and rare peloids occur (sample Flo-16). **C** – Wackestone–packstone. Peloids are the main components, along with benthic foraminifera (sample Bah-1). **D** – Wackestone–packstone. Peloids, ostrocodes (arrow) and benthic foraminifera (above the letter D) are the recognisable components (sample Bah-4).

E–H Fossil material: **E** – Wackestone. Components include peloids and miliolid foraminifera (arrow) (sample Re-17.11). **F** – Wackestone. Peloids are the main components, along with ostracodes (arrow) (sample Re-17.14). **G** – Slightly dolomitized wackestone including miliolids, *Pseudocyclammina* (arrow), and broken bivalve shells (sample Re-17.15). **H** – Wackestone with charophytes (c), peloids, coated intraclasts (arrow), and bivalves (sample Re-17.18).

Locality, sample number	Thickness of 20-ka sequence (decompacted) (cm)	Estimated duration of accumulation (a)	Major facies in sequence	Average accumulation rate (mm/a)
Col de l'Epine CE-28	40	6000 (but top eroded)	peloidal packstone, mudstone	0.07
Salève Sa-37	90	6000	peloidal packstone with oncoids	0.15
Salève Sa-97, Sa-98	175	6000	lagoonal packstone-wackestone	0.29
Reuchenette Re 17.9	75	6000	tidal flat and pond	0.13
Reuchenette Re 17.10, Re 17.11	140	20000	lagoonal wackestone	0.07
Reuchenette Re 17.12	75 ?	20000 ?	lagoonal mudstone	0.04 ?
Reuchenette Re 17.13	260	15000	lagoonal mudstone	0.17
Reuchenette Re 17.14, 17.15	140	6000	lagoonal wackestone	0.23
Reuchenette Re 17.16, 17.17	225	6000 (but top eroded)	tidal flat, lagoonal mudstone	0.38
Thickest sequences:				
Salève Sa-27 to Sa-33	730	20000	lagoonal packstone-wackestone	0.37
Reuchenette Re 23.16 to 23.20	1170	20000	lagoonal packstone-wackestone	0.59

Table 2 Estimated sediment accumulation rates for Lower Kimmeridgian and Lower Berriasian sequences. For interpretation see text.

facies is overprinted with dolomite. This indicates a lowering of sea level and conditions close to emersion. The following 20-ka sequence displays bird's eyes at its bottom and top. It is terminated by an erosion surface that is intepreted to represent a 100-ka sequence boundary (COLOMBIÉ, 2002). For the estimation of the corresponding sedimentation rates, it has been assumed that accumulation time of the 20-ka sequences with erosion or emersion features was 6000 years, whereas the full 20,000 years were allowed for the sequences that do not show signs of emersion or erosion (Table 2). In addition, the thickest sequences of the Reuchenette outcrop (COLOMBIÉ, 2002) and of the Salève outcrop (STRASSER & HILLGÄRTNER, 1998) have been chosen to estimate sedimentation rates when accommodation was at its maximum (Fig. 12).

The average accumulation rates for the studied sequences vary between 0.07 and 0.6 mm/a (Table 2). This is lower than in the Holocene examples. However, it has to be considered that large uncertainties are inherent in the estimation of decompaction factors and accumulation time.

5. DISCUSSION AND CONCLUSIONS

In Holocene shallow carbonate systems, it is seen that sediment accumulation rates vary laterally as well as through time (e.g., BOSENCE et al., 1985; CALHOUN et al., 2002; YANG et al., 2002; GISCHLER, 2003). These changes depend not only on sediment supply and accommodation, but also on local factors such as the pre-existing morphology of the sea floor that may influence the carbonate-producing organisms, and currents that redistribute the sediment. With Holocene sea-level rise flooding the platform, the sediments first record a deepening trend related to this transgression. However, with a decrease in the rate of sea-level rise, and sediment filling in the available space, a shallowing-upward facies evolution and progradation set in. In south-western Florida, the turn-around from retrogradation to progradation has been dated at about 3500 years BP (PARKINSON, 1989). On the isolated platforms in Belize, YANG et al. (2002) have shown that carbonate sediments accumulate slowly during early transgression (0.2–0.5 mm/a), faster during the late transgressive and early highstand phases (1.1–1.7 mm/a), and fastest during highstand conditions (2.4–4.6 mm/a).

The theoretical effects of basin morphology are shown in Fig. 14. If sea level does not drop below the platform edge, lowstand deposits develop in depressions on the platform, while the highs are exposed to erosion, cementation, and karstification. In basins isolated from the open ocean, fresh-water lakes can form. Rising sea-level then leads to initial flooding of the exposed land and to reworking of lowstand material. Morphological barriers serve as thresholds, and isolated basins may be flooded instantly when this threshold is passed. This will lead to abrupt facies changes that, however, need not be synchronous all over the platform. A syn-

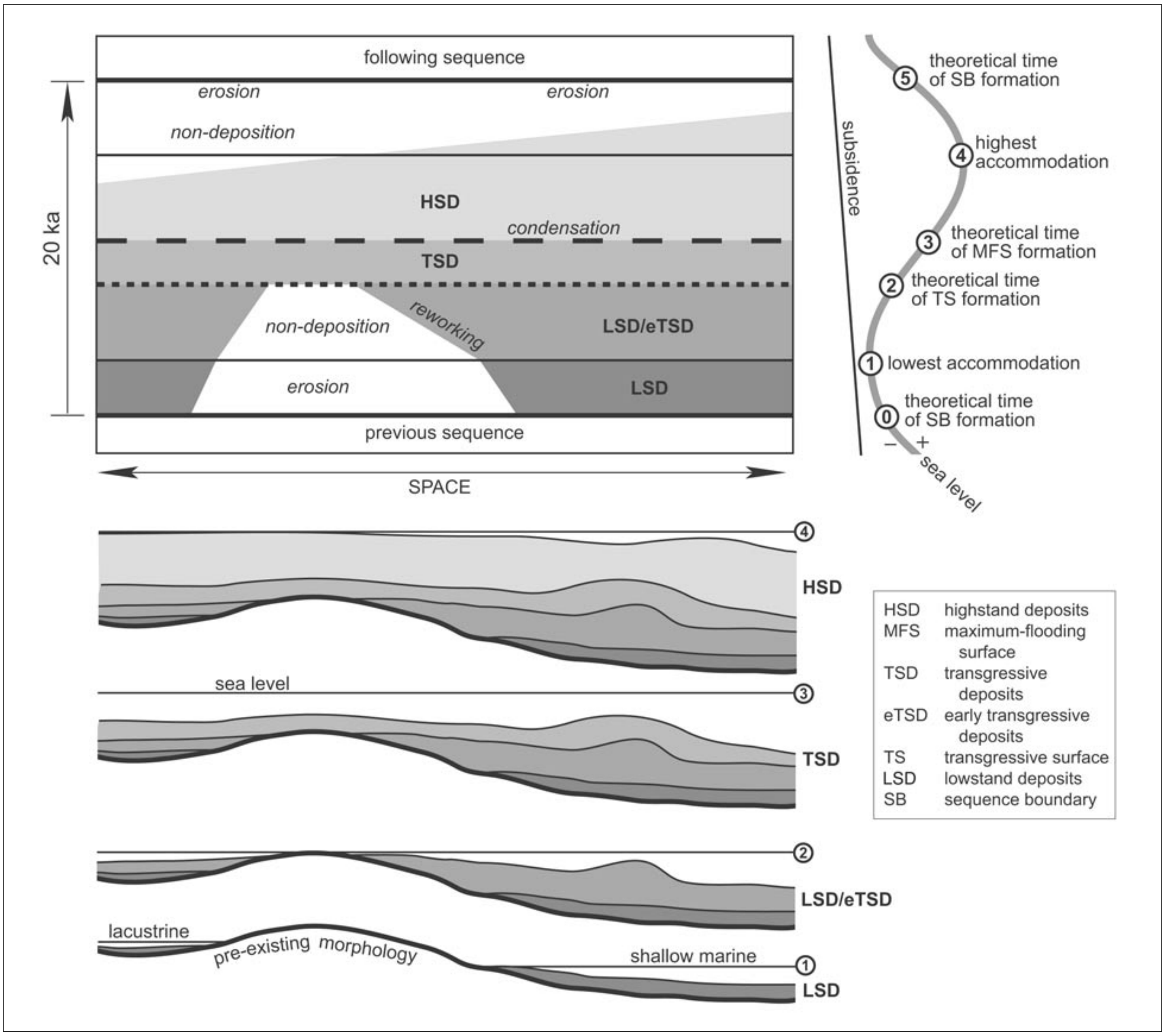

Fig. 14 Hypothetical space–time diagram illustrating the different durations of sediment accumulation depending on platform morphology. As a result of erosion, reworking, and condensation, sediment thicknesses may vary considerably through time and space.

chronous transgressive surface will only form once all islands are flooded. The final transgressive deposits will be relatively homogeneous all over the platform. During maximum flooding, condensation may set in if the water is too deep for efficient carbonate production. On the other hand, accommodation is increasing, and the thickest sediment package per time unit can potentially accumulate. With the lowering of sea-level, accommodation will be rapidly filled in, and the highstand deposits are forced to prograde. Further sea-level drop then exposes the sediment surface, and a sequence boundary forms. The consequence is that, during the same sea-level cycle, time available for sediment accumulation may vary dramatically from one place on the platform to the other.

There is no reason to believe that ancient platforms were less complex than modern ones, even if the high-frequency sea-level changes affecting them were of lower amplitude. To demonstrate this complexity, however, very high time resolution and lateral correlation of sections are needed. For example, STRASSER et al. (in print) interpret a step-wise flooding of the Jura platform in the Berriasian: each 20-ka sea-level pulse pushed ooid bars farther into the platform interior. SAMANKASSOU et al. (in print) show lateral facies variations within an Oxfordian 20-ka sequence over a few metres only: coral framestones are juxtaposed to ooid grainstones, and a tidal-flat facies is coeval with the top of a reef.

In the Holocene, the amount of carbonate produced on a shallow, tropical platform generally exceeds the amount of sediment accumulated on the platform top, and much of it is exported to the slope and the basin (DROXLER & SCHLAGER, 1985; MILLIMAN et al., 1993; SCHLAGER et al., 1994). Also in the Oxfordian of the Swiss and Swabian Jura there is evidence that carbonate was exported from the platform to the basin (PITTET et al., 2000). The rate at which sediment is produced (or supplied) thus does not correspond to the rate at which it accumulates in a given locality.

The Holocene sediments have accumulated on the platform tops over a relatively short time span. Sea level has not yet dropped, and most of the sediment that accumulated is also preserved (although it may be pushed back and forth on migrating islands and sand bodies). In the ancient sequences, however, erosion surfaces and vadose caps commonly indicate sea-level drops below the sediment surface, and part of the originally accumulated sediment may be missing. If long-term (million-year scale) accommodation gain was low or negative, the measured sediment thickness will of course not correspond at all to the accumulation potential of an individual sea-level cycle. Consequently, it would be erroneous to conclude on the health of a shallow-water carbonate system based only on the sediment that is preserved.

In Florida, on the Bahamas, and in Bermuda, accommodation gain over the last 6000 years was about 5 m (Fig. 10), and most Holocene sediment accumulated within this time span. The studied Kimmeridgian and Berriasian elementary sequences are, when decompacted, commonly a few metres thick (and may reach 7 or 11 metres; Table 2). Accommodation gain thus seems to be comparable between the Holocene and the ancient settings. For comparable lagoonal–peritidal facies, the Holocene sediment accumulation rates appear to be somewhat higher (0.3–3 mm/a; Table 1) than the ancient ones (0.07–0.6 mm/a; Table 2). However, the Kimmeridgian and Berriasian values are still higher than the ones published for similar ancient environments (0.03–0.09 mm/a; WILSON, 1975; SCHLAGER, 1981; ENOS, 1991).

The differences between Holocene and ancient accumulation rates as estimated in this study may be due to methodological errors, but can also imply that carbonate production on the Jura platform was lower during Kimmeridgian and Berriasian times. The reasons for this may be differences in water temperature and water chemistry, the generally higher clay input, and/or evolutionary changes in the carbonate-producing organisms. If production rates are assumed to have been similar to those in the Holocene, the lower net accumulation rates on the ancient platforms may result from increased carbonate mud export towards the basin (PITTET et al., 2000).

Although our study is punctational and only compares a few selected Holocene sequences with a few ancient ones, the following conclusions can be drawn:

– It is useful to distinguish between supply (or production) rates, accumulation rates, and preservation rates. Depending on the purpose of the study, the appropriate rate has to be estimated.
– Sediment accumulation rates must be estimated over a time span as short as possible in order to avoid averaging that also includes hiatuses or condensation.
– Considering the complexity of shallow-water carbonate platforms, it is dangerous to use values that average over large distances and across facies belts.
– There is a relatively good correspondence of estimated sediment accumulation rates between similar facies in the Holocene and Kimmeridgian or Berriasian carbonate systems. The somewhat lower rates for the ancient sediments may be due to methodological errors, to differences in the ecology of the carbonate-producing organisms, and/or to differences in sediment redistribution.

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