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Late Jurassic climate and its impact on carbon cycling

H. Weissert, H. Mohr

Geological Institute, ETH Zürich, CH-8092 Zürich, Switzerland

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Abstract

The climate of the Late Jurassic has been characterized by high atmospheric CO_2 levels and by a monsoonal rainfall pattern. In this study we traced the evolution of the global carbon cycle through the Late Jurassic with the help of carbonate carbon isotope stratigraphy. The Oxfordian–Tithonian $\delta^{13}C$ curve is marked by one major positive carbon isotope excursion with an amplitude $\Delta\delta^{13}C > 1.0‰$ (Middle to Late Oxfordian) and a second, minor positive excursion with an amplitude $\Delta\delta^{13}C < 1.0‰$ (Late Kimmeridgian). The Early Kimmeridgian and Early Tithonian $\delta^{13}C$-values fluctuate around $\delta^{13}C = +2‰ \pm 0.3‰$ and contrast with the less positive $\delta^{13}C$-values of Early Oxfordian and Late Tithonian age ($\delta^{13}C = +1‰ \pm 0.5‰$). A comparison of the Late Jurassic carbonate carbon isotope curve with the occurrence of organic-rich sediments suggests that not only fluctuations in organic carbon burial but also in carbonate carbon burial had an impact on the C-isotope record. The Oxfordian C-isotope excursion appears to correspond to a time of overall increased organic carbon burial triggered by increased nutrient transfer from continents to oceans during a time of rising global sea level. However, episodes of enhanced organic carbon burial during the Kimmeridgian and Early Tithonian are not reflected by prominent spikes in the C-isotope record. Favourable conditions for carbonate platform growth at a time of high global sealevel may have resulted in the stabilisation of the C org/C carb burial ratio and hence maintained the $\delta^{13}C$ record at steady but relatively positive values. The Middle and Late Tithonian C-isotope values drop below $\delta^{13}C = +1.5‰$. A similar shift to less positive C-isotope values was recognized in other C-isotope records from the Tethys and Atlantic Oceans and reflects a decrease in the C org/C carb burial ratio possibly related to a reorganisation of the global climate system.

Globally widespread marine black shale–mature quartzose sandstone assemblages suggest that the relative efficiency of the Late Jurassic carbon pumps was controlled by weathering, erosion and runoff causing widespread marine eutrophication. Eutrophication favoured the organic carbon pump but it diminished the carbonate-platform growth potential. The monsoonal rainfall distribution pattern may explain why Late Jurassic carbonate platforms experienced less severe growth crises than Early Cretaceous carbonate platforms.

1. Introduction

"The function of life, considered from the material point of view, is not only fundamentally concerned with the atmosphere, and intimately dependent on its conditions, but its most important material effects appear to lie in its modification of the constitution of the atmosphere".

(*Chamberlin and Salisbury, p. 644, 1906*)

Almost 60% of all the presently known fossil fuel reserves are derived from source rocks of Late Jurassic and middle Cretaceous age (Ulmishek and Klemme, 1990). The deposits enriched in marine and terrestrial organic carbon were formed in shallow marine settings, on open shelves or even in deep sea environments. The occurrence of organic-rich sediments of both Late Jurassic and

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middle Cretaceous age is commonly linked to increased productivity and/or low oxygen/anoxic bottom water and/or high burial rates of organic carbon favouring preservation of organic carbon (Schlanger and Jenkyns, 1976; Weissert et al., 1979; Arthur, 1982; Bralower and Thierstein, 1984; Calvert, 1986; Tyson, 1987, and many others). The Late Jurassic was—in contrast to the Cretaceous—not only a time of increased organic carbon burial but it was also a time which was favourable for reef growth and for increased accumulation of carbonate carbon (Budyko et al., 1987).

In this study we shall document how the Late Jurassic carbonate carbon isotope record mirrors the response of the marine organic and carbonate carbon pumps (Volk and Hoffert, 1985) to changing climate. The excessive burial of organic carbon at times of high carbonate accumulation rates was triggered by a greenhouse climate combined with a monsoonal rainfall pattern controlled by the break-up of the Pangea supercontinent during the Late Jurassic (Sheridan, 1983; Ziegler, 1988; Parrish, 1993). Intensified seafloor spreading and volcanic activity impacted Jurassic climate through the increased flux of volcanic CO_2 to the atmosphere, as suggested by a number of GCM-paleoclimate models (Moore et al., 1992; Valdes and Sellwood, 1992) and by geochemical models (Berner, 1994).

The history of atmospheric CO_2 levels and of the global carbon cycle is mirrored in marine carbonate carbon isotope curves (Scholle and Arthur, 1980). Fluctuations in $\delta^{13}C$ can be explained by variable C org/C carb export ratios into the marine sedimentary carbon reservoir (Schidlowski, 1987). Times of intensified biological carbon pumping and of high C org burial rates are commonly linked to perturbations of the atmospheric carbon reservoir and to greenhouse climate. Positive $\delta^{13}C$-excursions may be regarded as response signals of perturbation of global climate linked to fluctuations in atmospheric carbon dioxide concentrations (Arthur et al., 1985; Weissert, 1989). The limestone archives which were formed along the margins of the opening Late Jurassic Tethys Ocean offer the opportunity to trace long term changes in the C-isotope record.

While the Mesozoic C-isotope curve records episodic perturbations of Mesozoic carbon cycle and climate, other tracers of paleoenvironmental change should contain signatures of monsoonal rainfall pattern. Accelerated water cycling, which according to climate models coincides with high energy or greenhouse episodes (Del Genio et al., 1991), should have left an imprint in the accumulation history of continental clastics in marine archives. We expect the distribution of Late Jurassic siliciclastics within the framework of an increasingly fragmented Pangea continent to reflect paleo-rainfall conditions.

2. Carbon isotope stratigraphy

An extensive Late Jurassic–Cretaceous C-isotope stratigraphy has already been obtained from pelagic sediments of the southern continental margin of the Alpine Tethys Ocean now outcropping in the Southern Alps (Weissert et al., 1985; Weissert and Channell, 1989; Weissert and Lini, 1991; Lini et al., 1992). Weissert and Channell (1989) documented how the Late Jurassic carbonate carbon isotope curve shifts from relatively positive $\delta^{13}C$-values near $+3.0‰$ in the Kimmeridgian–Early Tithonian to values near 1.30‰ in the Late Tithonian–Early Berriasian. Weissert and Channell (1989) dated the transition to low $\delta^{13}C$-values in the latest Jurassic to occur within Magnetozones M19–M21.

For this study we have selected pelagic sediments deposited on the shallow northern shelf of the Alpine Tethys Ocean for additional carbon isotope measurements. We wanted to test the reproducibility of the Southern Alpine C-isotope record and to extend it back through the Oxfordian. The Northern Tethyan sediments offer a further advantage of linking the C-isotope record with a sea level history.

Bulk samples were chosen for the establishment of a C-isotope stratigraphy of the Mid-Oxfordian Schilt Formation and of the Late Oxfordian–Tithonian Quinten Limestone Formation. The samples were prepared and analyzed following standard preparation techniques (Mc Crea, 1950). A VG-Micromass 903 mass-spectrometer provided data with a reproducibility of $\pm 0.2‰$ for duplicate sample analyses and ±0.01‰ for replicate standards.

3. Northern Tethyan limestones: an archive for a Late Jurassic C-isotope record

The sedimentary rocks carrying the Oxfordian–Tithonian C-isotope signal are preserved in the Helvetic nappes of eastern Switzerland (Fig. 1). The base of the sequences investigated consists of micritic and partly dolomitized limestones and marls of the Schilt Formation (Fig. 2). Condensed nodular limestones (Schilt Limestone Member) were deposited on shallow current-dominated environments of the northern Tethyan shelf while marls of the Schilt Formation were deposited in deeper settings less affected by current activity (Schilt Marl Member). The Schilt limestones are rich in siliceous sponges and ammonites and reach a thickness of up to 20 m (Kugler, 1986). The less condensed Schilt marls reach a thickness of up to 60 m. The base of the Quinten Formation consists of dark micritic limestones with a carbonate content of up to 95% and organic carbon contents reaching up to 1% and rare marl interbeds. It may reach a thickness of more than 400 m. This formation is grouped into three members: The Lower Quinten Limestone, the Marlstone Member and the Upper Quinten Limestone. The Lower Quinten Limestone is built up by regulary bedded micritic limestone containing radiolarians, rare sponge spicules, and crinoidal debris. An increase in fine carbonate clastics in the upper part of the Lower Quinten Limestone can be interpreted as an indicator of a shallowing upward trend. The overlying Marlstone Member consists of thin bedded micritic and partly dolomitized limestones. The Upper Quinten Limestone Member shows a less regular bedding of the micritic limestone that is, in its most proximal environments, gradually replaced by a prograding carbonate platform with coral reefs known as the Troskalk Formation.

The nodular Schilt limestones and the Schilt marls contain an ammonite fauna of the *transversarium*-zone (Middle Oxfordian, Kugler, 1986) The transition of the Schilt Limestone to the Lower Quinten Limestone falls into the *bimammatum* zone of Late Oxfordian (Kugler, 1986). The Lower Quinten Limestone Member remains poorly dated and covers part of the Kimmeridgian. Ammonites in the Marlstone Member were assigned to the *pseudomutabilis* Zone of the Late Kimmeridgian (Kugler, 1986). The Upper Quinten Limestone

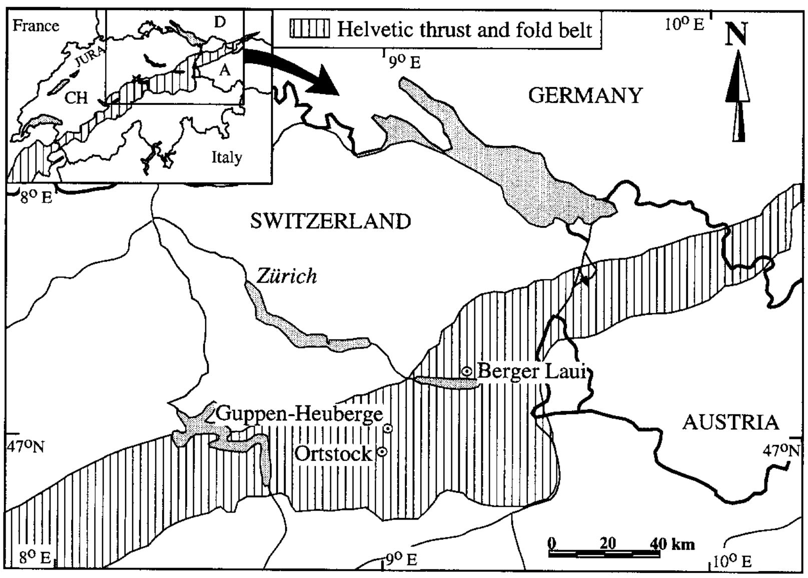

Fig. 1. Location map showing the sections described in the text and illustrated in Fig. 2.

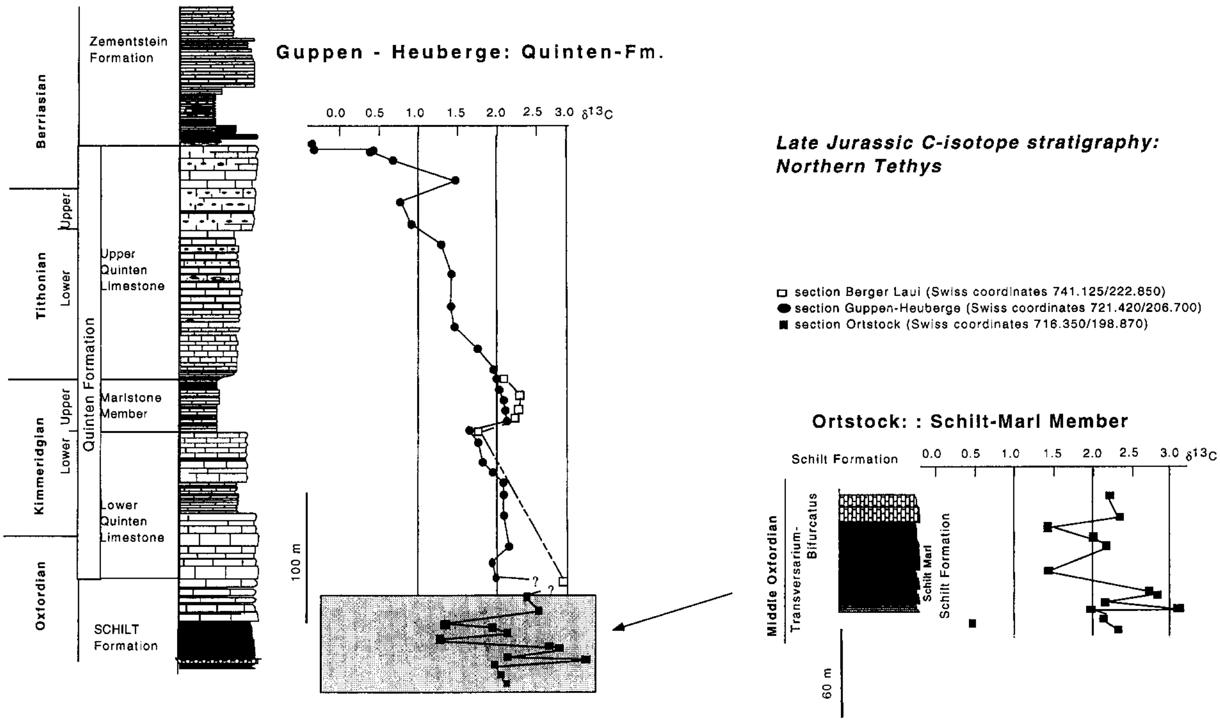

Fig. 2. Litho-, bio- and carbon isotope stratigraphy of the Quinten Limestone Formation and the Schilt Formation in northern Switzerland. Ortstock section and data are shown at the base of the Guppen-Heuberge section to facilitate comparison of $\delta^{13}C$-data. A detailed description of these sections is given by Kugler (1986) and Mohr (1992).

was deposited during the Tithonian. Its uppermost part, which is affected by the southward progradation of a carbonate platform ("Troskalk"), is dated with calpionellids and falls within the Late Tithonian calpionellid zone B.

4. A northern Tethyan carbon isotope record

The C-isotope data are graphically summarized in Fig. 2. The measured carbon isotope record starts with a positive shift starting within the Oxfordian *transversarium* zone. In the section "Ortstock" we measured a C-isotope shift of more than 1‰ in the basal part of the Schilt Formation. The most positive Oxfordian values fall within a range of +2.9‰ to +3.1‰ which corresponds to a synchronous carbon isotope excursion measured in samples from the Swiss Jura mountains (Bill et al., 1995; see Fig. 4). One very positive carbon isotope value (+2.95‰) was measured at the base of the Quinten Limestone Formation in the section Berger Laui. Otherwise, the Latest Oxfordian and Early Kimmeridgian C-isotope values of the Quinten Limestone Formation fluctuate around +2.3‰. C-isotope values near $\delta^{13}C = +1.8$‰ mark the base of the Upper Kimmeridgian. A second minor positive C-isotope excursion with an amplitude of $\Delta\delta^{13}C = 0.5$‰ is dated as Late Kimmeridgian–Early Tithonian in age. The Tithonian is marked by a gradual decrease in C-isotope values. A difference $\Delta\delta^{13}C = -1.0$‰ recognized between Early and early Late Tithonian samples corresponds to a comparable decrease in carbonate carbon isotope records established in the Southern Alps and in the North Atlantic (Brennecke, 1977; Létolle et al., 1978; Weissert and Channell, 1989).

The oxygen isotope values fluctuate between −3‰ and −6‰ in the limestones from the section "Guppen-Heuberge" and "Berger Laui". The

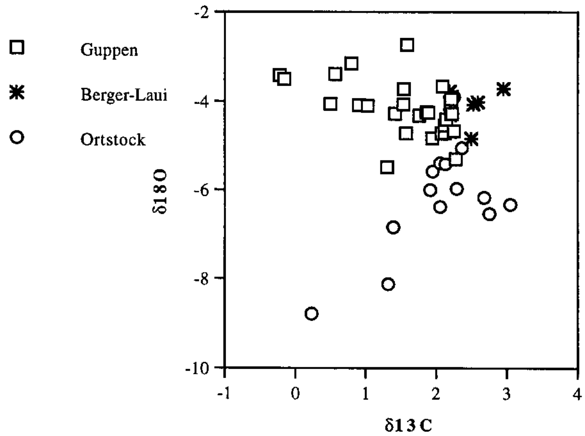

Fig. 3. Oxygen-carbon isotope plot of analysed samples from the sections Guppen-Heuberge and Ortstock in northern Switzerland. (Tables with the complete data set are available from the first author).

slightly more metamorphosed samples from the "Ortstock" section (Schilt marls) show oxygen isotope values between −5.3‰ and −8.8‰ (Fig. 3). The values are interpreted as a result of limestone recristallisation during low grade Alpine metamorphism which affected the sections studied. Similar trends to more negative oxygen isotope values in recristallized Alpine Jurassic limestones were observed in a study by Green-Früh et al. (1991). Low grade Alpine metamorphism and isotopic homogenisation may have masked evidence of meteoric diagenesis near the sequence boundary separating the Quinten Formation from the overlying Cretaceous carbonate and marl sequences. While the oxygen isotope data did not preserve a distinct meteoric signature, the large Late Tithonian C-isotope shift reaching values of $\delta^{13}C = -0.2‰$ may contain a meteoric diagenetic component which was preserved during alpine recristallisation. Therefore the C-isotope data from the top of the Quinten Formation are not included into the discussion on Late Jurassic carbon cycling.

5. Carbon isotopes and the carbon budget

Late Jurassic C-isotope stratigraphies established in other paleoceanographic settings match the pattern recognized in northern Tethyan sediments. Pisera et al. (1992) analyzed bulk carbonate samples and brachiopod shell material from a Late Jurassic section in Poland and Germany. They identified a positive C-isotope excursion of Late Oxfordian age. Dean and Arthur (1987) analysed the C-isotopic composition of bulk organic carbon from Atlantic drill sites of the "COST" project. Their data show a remarkable transition of Late Jurassic C-isotope values fluctuating around $\partial^{13}C = -22‰$ to more negative values near $\partial^{13}C = -24‰$ at the end of the Tithonian. Heydari et al. (1993) anlysed Late Oxfordian sediments from the Smackover Formation and they report C−isotope compositions reaching values of up to $\delta^{13}C = +3‰$ to $+6‰$. Jenkyns (in press) identified a prominent C-isotope excursion falling within the *transversarium* zone and reaching carbon isotope values of more than $+3‰$ in sections from the southern Tethyan margin.

The observed variations in the Late Jurassic carbonate carbon isotope record may reflect changes in the Jurassic marine carbon cycle. Times marked with more positive C-isotope compositions should correspond to times of a higher organic carbonate/carbon burial ratio (Schidlowski, 1987). The long term oceanic carbon isotope budget should reach a balance if average river carbon input is balanced by carbon and carbonate export into the sedimentary sink:

$$100(\delta^{13}C)_{in} = (100 - R)\delta^{13}C\ cc_{out} + R\delta^{13}C\ org_{out}$$

$(\delta^{13}C)_{in} = -6‰$, $\delta^{13}C\ cc_{out} = +1.25–+2.9‰$, $\delta^{13}C\ org_{out} = -24––22‰$

If we use average values for carbon entering the ocean in dissolved form ($\delta^{13}C_{in} = -6‰$) and for organic carbon being deposited in marine sediments ($\delta^{13}C$ org $_{out} = -24––22‰$), we may explain an average amplitude $\Delta\delta^{13}C$ cc $= 1.5‰$ in the Late Jurassic carbonate carbon isotope record with a 12% up to 20% variation in average global marine C org accumulation rates assuming constant carbonate carbon burial rates. If this assumption is correct, then the Middle to Late Oxfordian with its most positive C-isotope values should have been marked by highest accumulation rates of organic carbon. Another interval of increased accumulation of organic carbon should fall within the Late Kimmeridgian.

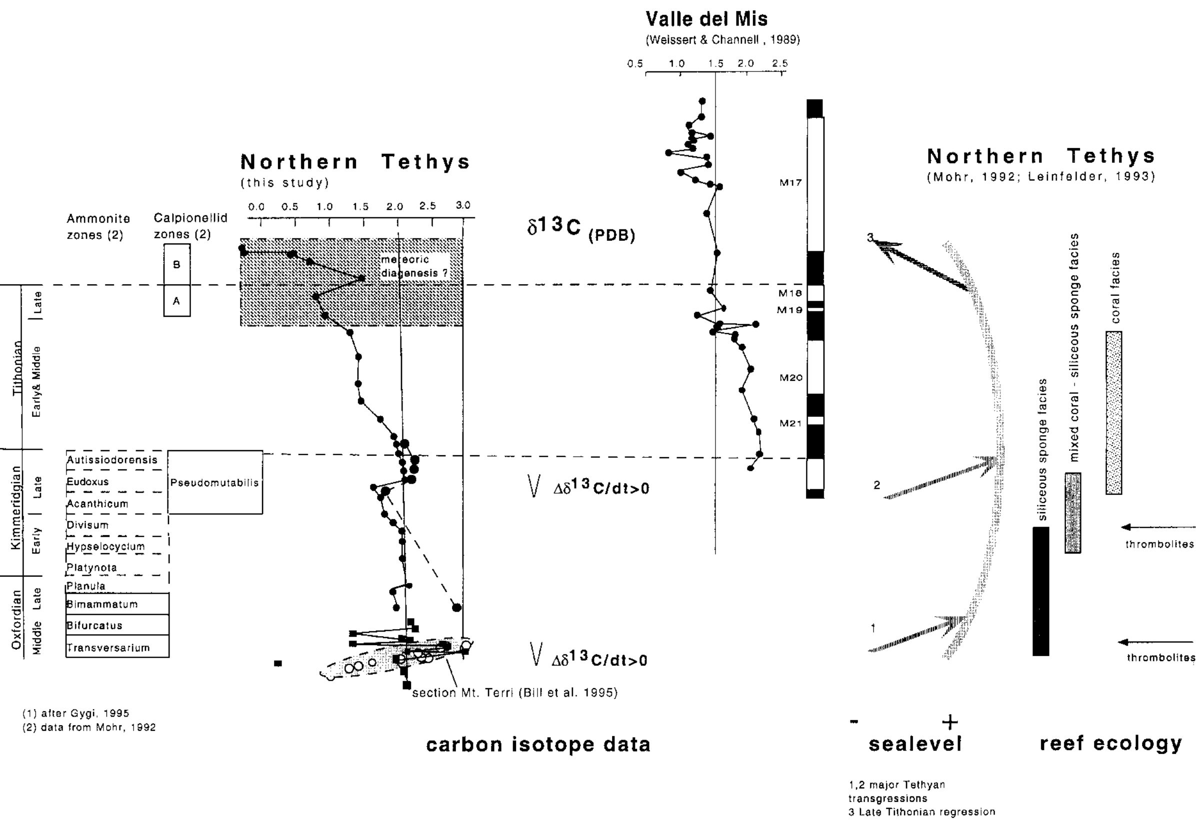

Fig. 4. A comparison of the measured C-isotope curve with C-isotope records from the Swiss Jura Mountains (Bill et al., 1995) and the Southern Alps (Weissert and Channell, 1989), with trends in the Late Jurassic sealevel curve (Mohr, 1992; Haq et al., 1987) and with the changes in northern Tethyan reef ecology (Leinfelder, 1993).

Available data on organic carbon accumulation rates do indicate that marine or terrestrial organic carbon was deposited at elevated rates into a variety of Late Jurassic marine depositional environments. Today, Late Jurassic deposits form the source for up to 29% of available fossil fuel reserves (Ulmishek and Klemme, 1990).

A more accurate survey of the stratigraphic position of major organic carbon-rich deposits shows major peaks in the Late Oxfordian, the Kimmeridgian and in the Early Tithonian (Fig. 5). Major sinks for organic-rich sediments were the cratonic Siberian Basin (Oxfordian and Volgian/Tithonian age, Nesterov et al., 1990), the narrow North Sea (Oxfordian–Kimmeridgian, Brown, 1984), the North American Gulf Coast (Smackover Formation, Late Oxfordian; Claypool and Mancini, 1989; Fails, 1990), and the Arabian shield, where organic-rich sediments accumulated in intrashelf basins. These organic carbon-rich sediments of the Arabian shield form the world's richest single oil deposit (Oxfordian–Kimmeridgian, Alsharan, 1993).

The widespread occurrence of organic carbon-rich sediments can be explained as a result of high productivity and/or limited deep-water renewal and widespread low oxygen conditions and/or elevated sedimentation rates with improved preservation of marine and terrestrial organic carbon (e.g. Tyson, 1987). Examples for C org-rich sediments related to high productivity are given by the Kimmeridge clay (e.g., Tribovillard et al., 1992) or by the Siberian Bazhenov suite, which is of Volgian age (e.g., Nesterov and Ushatinsky, 1991).

Table 1
Oxygen and carbon isotope data

(m)	$\delta^{13}C$	$\delta^{18}O$
Section Guppen-Heuberge		
1	1.304	−5.492
13	2.123	−4.537
30	2.08	−4.715
40	2.283	−5.301
70	2.207	−4.086
90	2.215	−3.951
100	2.203	−4.053
120	1.941	−4.827
140	1.89	−4.257
150	1.761	−4.318
160	2.252	−4.661
170	2.225	−4.276
180	2.214	−4.276
190	2.143	−4.405
200	2.122	−4.705
210	2.094	−3.656
230	1.865	−4.245
250	1.574	−4.721
270	1.532	−4.069
300	1.542	−3.721
330	1.418	−4.277
350	1.03	−4.1
370	0.901	−4.082
390	1.587	−2.723
410	0.795	−3.142
419	0.497	−4.05
420	0.565	−3.377
421	−0.156	−3.493
426	−0.222	−3.412
Section Berger Laui		
30	2.95	−3.71
148	2.2	−4.68
150	2.5	−4.84
155	2.59	−4.01
170	2.53	−4.07
175	2.53	−4.07
205	2.22	−3.89
210	2.2	−4.31
212	2.19	−3.77
Section Ortstock		
0.1	2.29	−5.97
0.5	0.23	−8.8
1	2.13	−5.41
3	1.92	−6
5	3.05	−6.33
5.5	2.06	−6.38
12	2.76	−6.54
14	2.68	−6.17
22	1.32	−8.13
44	2.06	−5.39
48	1.95	−5.58
54	1.39	−6.85
66	2.37	−5.05
88	2.25	−3.9

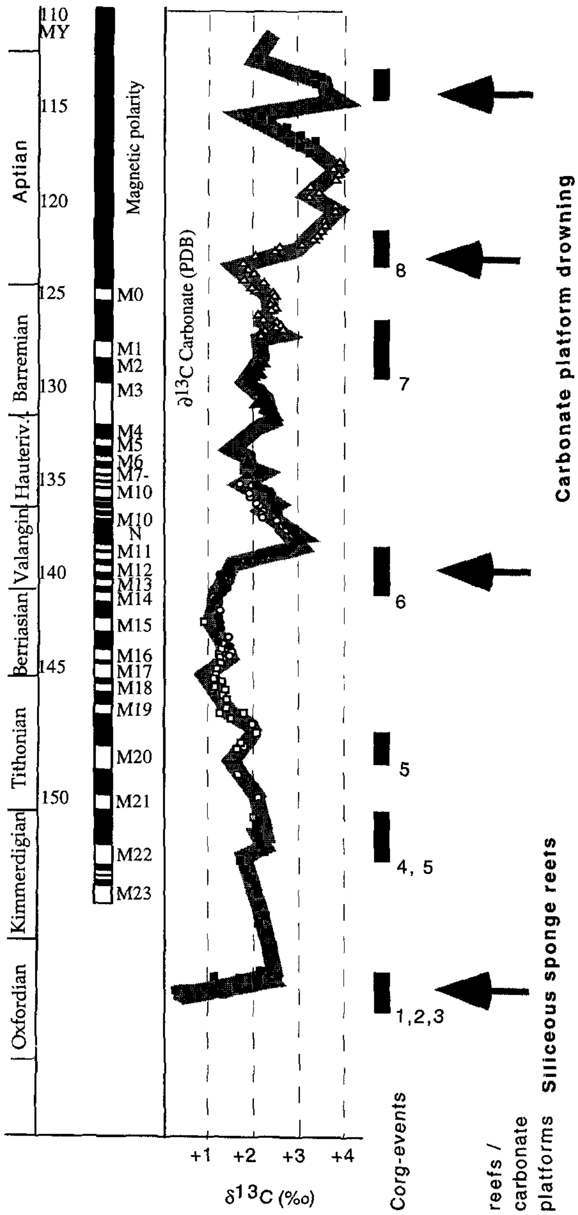

Fig. 5. Composite carbon isotope stratigraphy with data from the southern Alps (Weissert et al., 1985; Weissert and Lini, 1991; Lini et al., 1992) and additional data from this study. Major episodes of accumulation of organic carbon in different paleoenvironmental settings are marked with black bars: *1* = Siberia, *2* = Smackover, *3* = Arabian shield, *4* = Siberia (Bazhenov), *5* = North Sea, *6* = North Atlantic, *7* = OAE I (Selli), *8* = OAE II (N. Jacob). References see text. Evolution of reef and carbonate platform ecology after Leinfelder (1993) and Föllmi et al. (1994).

As one example of a circulation-controlled accumulation of excessive organic carbon, we may mention the Late Jurassic oil source rocks in the North Sea. There, Oschmann (1988) considers restricted deep water circulation related to monsoonal climate as the major cause of accelerated organic carbon burial.

Thus it seems that paleoceanographic and paleoenvironmental conditions favoured the accelerated burial of organic carbon not only during the Oxfordian but also at Kimmeridgian and Early Tithonian times. While the Oxfordian episode of enhanced organic-carbon burial seems to have left a distinct signature in the C-isotope record we see only a minor C-isotope excursion coinciding with the episodes of high organic carbon burial during the Kimmeridgian and the Early Tithonian ($\Delta\delta^{13}C = 0.5‰$). The Kimmeridgian and Early Tithonian C-isotope curve seems to have stabilised at relatively positive values, which suggests that elevated organic carbon burial rates were buffered by the Late Jurassic carbonate carbon pump.

6. Climate, weathering and the organic carbon pump

The carbon isotope record and its preliminary interpretation as a proxy for organic carbon burial provokes two questions: (1) In the case that biological carbon pumping was intensified over a period of up to millions of years, the limiting nutrients phosphate and nitrate also had to be provided at an increased rate (Weissert, 1989). (2) If intensified biological carbon pumping is seen as a response signal to elevated atmospheric carbon dioxide levels (Weissert and Lini, 1991), then other elements of the biosphere should reflect environmental conditions marked by elevated atmospheric energy. In the following we will document how altered atmospheric energy conditons triggered the intensification of global water cycling leading to accelerated nutrient transfer from continents to oceans.

Siliciclastics deposited at an accelerated rate in tectonically non-convergent oceanic settings can serve as a source of information on the weathering intensity and therefore point at the mode of nutrient cycling (Cecil, 1990; Weissert, 1990). Chemically mature sands today form in low relief regions characterized by transport-limited erosion and by warm and humid climate. Warm-humid fluvial environments also provide time for chemical weathering of orogenically derived immature sands during extended fluvial sand storage in alluvial plains (Johnsson et al., 1991). Paleoclimate maps for the Late Jurassic allow us to identify potential areas of increased net precipitation favouring biogeochemical weathering (Fig. 6; Moore et al., 1992). If greenhouse climate episodes were coupled with elevated weathering rates, then we should find their signatures in oceanic sinks off the regions of high net precipitation and transport-limited erosion. Indeed, at several of these localities we find a remarkable association of Upper Jurassic black shales recording conditions favourable for excessive organic-carbon burial and interbedded mineralogically mature siliciclastics which we interpret as signatures of intensified chemical weathering (Fig. 6).

A major rainfall belt along N-America–Greenland can explain the occurrence of mainly quartzose sandstones intercalated with black carbonaceous mudstones in eastern Greenland (Hareelv Formation, Surlyk, 1987). Medium to coarse grained glauconitic quartz sandstones were also shed into the Middle to Late Oxfordian–Kimmeridgian North Sea. The carbonaceous mudstones and quartz-sandstones today are known as one of the important oil sources and reservoirs in the North Sea region (Harker et al., 1993). The kaolinite distribution pattern in the Kimmeridge Clay of France and England is interpreted as reflecting humid conditions during the Late Jurassic that contrasts with increasing aridity moving towards the earliest Cretaceous (Hallam et al., 1991; Wignall and Ruffell, 1990). In southern France, kaolinite occurrences in Kimmeridgian and Lower Tithonian hemipelagic marls from the Vocontian Trough again reflect strongly humid conditions during the Late Jurassic in low latitude northern Tethyan areas (Deconinck et al., 1985). Humid climate also affected the carbonate platforms along the northern margin of the Alpine Tethys Ocean as evidenced by the widespread occurrence of kaolinite in Oxfordian–

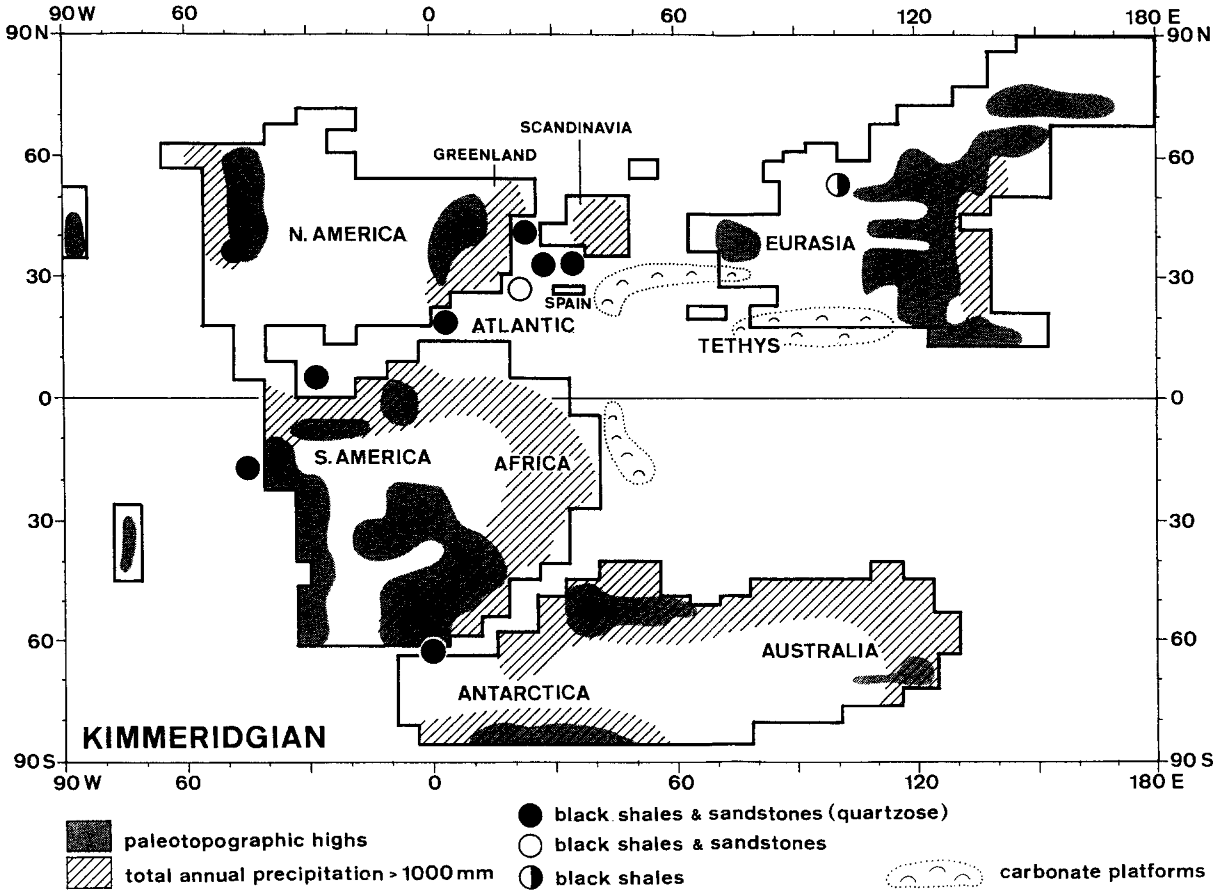

Fig. 6. Precipitation pattern in the Late Jurassic according to palaeoclimate models by Moore et al. (1992). Areas of black shale–sandstone/quartz sandstone accumulation are plotted on the map. In addition, areas of widespread shallow water carbonate accumulation are shown (references in text).

Kimmeridgian sediments outcropping in the Swiss Jura mountains (Gygi and Persoz, 1987).

Thick sandstones intercalated with organic-rich sediments were deposited along the Grand Banks off North America (Sinclair, 1988). Upper Jurassic sediments today outcropping in Mexico formed along the Carribbean margin. They are rich in silt-sized quartz and in kaolinite. Salvador et al. (1992) relate the inflow of fine terrigenous clastics into the Kimmeridgian–Tithonian Gulf of Mexico to a climate with moderate to high rainfall.

Mature sandstones were also deposited near the mid-latitude rainfall area in western North America. There, black shales of the Upper Jurassic Fernie Formation with low organic carbon contents alternate with turbiditic quartz sandstones (pers. obs.; Poulton et al., 1990). In Peru, the Upper Jurassic Oyon Formation consists of carbonaceous siltstones and shales interbedded with sandstones and whitish quartzites (Riccardi et al., 1992). Thompson (1976) describes a Middle to Upper Jurassic sequence of carbonaceous claystones with subarkose sandstone intercalations from DSDP drill sites on the Falkland Plateau. He noted frequent sandstone intercalations in Late Jurassic sediments contrasting with rare sandstone beds in the sediments marking the Jurassic–Cretaceous transition.

Nesterov and Ushatinsky (1991) speculate about the cause of elevated Late Jurassic marine productivity in the Siberian Basin and they conclude that intense continental weathering resulted in increased nutrient transfer rates and eutrophication of the Siberian Basin.

In view of the poorly constrained chronostratigraphy for the Late Jurassic, well defined accumulation rates for siliciclastics remain rare. Few data from early DSDP Sites suggest, that bulk sediment accumulation rates in the Kimmeridgian were significantly higher than at the Jurassic–Cretaceous transition (Thiede and Ehrmann, 1986). If these increased sediment accumulation rates can be construed as evidence for intensified climate induced weathering and erosion remains debatable.

The global seawater Sr-isotope curve is marked by a pronounced minimum in the Callovian–Oxfordian followed by a transition to ^{87}Sr enriched values is through the Late Jurassic (Jones et al., 1994; Fig. 6). Although changing hydrothermal inputs cannot be ruled out, the Sr-isotope record is consistent with increasing weathering rates throughout Late Jurassic and Early Cretaceous. Another potential indicator of runoff and weathering is the Nd-isotope record (Fig. 6; Stille and Chaudhuri, 1993). Deposits from the Tethys Ocean seem to preserve a Nd-isotope signature which strongly differs from the Pacific record and which reflects limited isotopic homogenization in Late Jurassic–Cretaceous oceans. Depleted Nd-isotope values in Tethyan sediments suggest that increased continental erosion and runoff affected the Nd-isotope signature of Tethyan–Atlantic seawater starting around 160 million years ago and reaching its most depleted values towards the end of the Jurassic and in the Early and Middle Cretaceous (Stille and Chaudhuri, 1993).

The combined sedimentological and geochemical evidence allows us to agree with Hallam (1984), who described the Late Jurassic as an exceptionally humid time in Mesozoic climate history. While the intense humidity corresponded to a time of elevated energy level of the Late Jurassic biosphere, the distribution of rainfall was controlled by a monsoonal climate which was progressively replaced by a more zonal paleoclimate at the turn into the Cretaceous. We conclude that specific weathering, erosion and runoff conditions facilitated the globally widespread accelerated organic carbon pumping in Late Jurassic oceans. If we take the carbonate C-isotope curve as a proxy for enhanced organic carbon burial we are confronted with the question why the carbon-isotope excursions in the Kimmeridgian–Tithonian C-isotope stratigraphy are far less prominent than in the Oxfordian, despite of episodic high organic-carbon burial rates during these times. Does the Late Jurassic sedimentary record also provide information on changing carbonate-carbon accumulation rates? Do we have evidence for decreased efficiency of the global carbonate carbon pump at the time of the pronounced positive C-isotope excursion in the Oxfordian? A carbon isotope excursion with an amplitude $\Delta\delta^{13}C = 1.5‰$ could have resulted from a decrease in global calcite sedimentation rates on the order of 10% at constant organic carbon burial rates. And, do we find evidence for crises in Middle to Late Oxfordian carbonate platform growth and for increased carbonate carbon burial rates during Kimmeridigan and Tithonian times?

Conditions favouring carbonate sedimentation include (1) a high global sealevel providing space for the aggradation and progradation of platform carbonates and (2) oligo- to mesotrophic neritic environments situated in low latitudes not affected by shedding of siliciclastics. High suspension and nutrient loads in coastal waters diminish the growth potential of carbonate buildups.

7. Sea level and the carbonate carbon pump

The Northern Tethyan archive for a carbonate carbon isotope record contains information on the relative sea level history during the Late Jurassic. Kugler (1986) interprets the middle Oxfordian Schilt limestones and marls as transgressive facies related to a sea level rise. He could correlate the data collected from the northern Tethyan shelf with information gained on the carbonate platform of the Swiss Jura mountains, situated northwest of the Helvetic shelf. Gygi (1986) studied the sedimentology and Late Jurassic facies pattern of the Jura mountains and he also found evidence for a major mid-Oxfordian transgression starting within the *transversarium* ammonite zone and peaking during Late Oxfordian and Early Kimmeridgian. The Quinten Marlstone Member with ammonites of Late Kimmeridgian age is interpreted as a lowstand deposit (Mohr, 1992). A renewed transgression resulted in a maximum flooding and a highstand systems tract which is of Tithonian age. The Late Tithonian shallowing upward sequence of the Quinten Limestone Formation with the overlying prograding reefoidal shallow water limestone is capped by a sequence boundary related to a regionally widespread drop in sea level.

Ponsot and Vail (1992) studied the Late Jurassic sealevel history of the Paris–London Basin. They also noted a mid-Oxfordian transgression (*bifurcatus* zone) marking the onset of a widespread Late Jurassic sea level highstand interrupted by several third order sealevel fluctuations (transgressive systems tracts: Late Oxfordian, Middle and Late Kimmeridgian). Leinfelder (1993) studied the Oxfordian to Tithonian carbonate shelf evolution along the northern Tethys from Germany to Portugal. The Late Jurassic sea level highstand was reached after a major mid-Oxfordian transgression (*transversarium* zone). In addition, he recognized transgressive facies in sediments of Late Oxfordian (*bimammatum* zone) and the Middle Kimmeridgian (*divisum* zone) age.

Transgressive conditions for the Oxfordian are reported not only from the Tethys Ocean but from areas as the Siberian Basin (Nesterov and Ushatinsky, 1991, and pers. obs.), or the Late Jurassic Gulf coast environment (Claypool and Mancini, 1989). Evidence for an Oxfordian sea level rise with a maximum flooding episode in the Kimmeridgian is reported from Greenland (Surlyk, 1987, 1990). Surlyk (1990) notes that thick sandstones were accumulated during sea level rise. An increase of fine clastics along the coast of the Gulf of Mexico during a relative sea level highstand at Oxfordian–Kimmeridigian time is described by Salvador et al. (1992). In Western Canada the accumulation of quartzose sandstones within the Late Jurassic upper Fernie Formation occurred during a major transgressive episode (Poulton, 1990; and pers. obs.).

If we compare the Late Jurassic sea level curve with the carbonate carbon isotope record we recognize that the onset of the Late Jurassic carbon isotope excursion coincides with a marine transgression while the drop at the end of the Tithonian correlates with the end of the Late Jurassic sea level highstand and the subsequent sea level drop. Analogous correlations between sea level and carbon isotope records are known from studies on the Valanginian and Aptian C-isotope excursions (Weissert and Lini, 1991; Föllmi et al., 1994) or on the Toarcian carbon isotope event (Myers and Wignall, 1987).

Based on the available information, the widespread inundated shelves provided new space for the accommodation of carbonate sediments in shallow water seas during the Late Jurassic (see also Hallam, 1992). Major reef provinces and carbonate platforms of Late Jurassic age include the European realm (Wilson, 1975; Gygi, 1986; Leinfelder et al., 1993 and many others), the Arabian shield and Central Asia (Wilson, 1975; Nalivkin, 1973). These reef provinces reflect the overall favourable conditions for carbonate buildup development during the Late Jurassic and they could explain why the high organic carbon burial rates could have been balanced by accelerated carbonate puming as recorded in the low amplitude C-isotope curve of the Kimmeridgian and Early Tithonian.

8. Weathering and carbonate platform growth potential

While inundated shelves provided conditions favourable for carbonate platform growth, the climate induced weathering regime resulting in increased transfer rates of siliciclastics and in eutrophication of Jurassic oceans limited the growth of carbonate buildups during the Middle and Late Oxfordian and to a certain extent during the Kimmeridgian. Environmental conditions favoured the growth of sponge reefs along the middle Oxfordian to Early Kimmeridgian margin of the northern Tethys (Leinfelder, 1993). A mixed coral reef–sponge reef facies established during the Kimeridgian reflects paleoceanographic conditions which were increasingly favourable for carbonate reef building organisms. Northern Tethyan sponge reefs were completely replaced by coral reefs in the Tithonian. Additional evidence for environmentally induced crises in reef ecology is given by the repeated occurrence of thrombolites in the Late Jurassic shelf and platform stratigraphy of the northern Tethys. Leinfelder (1993) placed episodes of widespread thrombolite formation and reef growth crises into the Middle and Late Oxfordian and the middle Kimmeridgian. Dysaerobic bivalves and abundant glauconite co-ocurring with the thrombolites provide evidence for low-oxgen water masses which we relate to widespread eutrophication of Oxfordian and Kimmeridigian water-masses. Leinfelder et al. (1993) further describe the environment which favoured the growth of the thrombolites as "contaminated by fine siliciclastics material".

We conclude that weathering and erosion were responsible for widespread eutrophication of Oxfordian coastal waters. The large Middle to Late Oxfordian C-isotope excursion may therefore carry a signature of increased organic carbon burial and of regionally diminished growth potential of carbonate platforms. The Kimmerigian and Tithonian reef growth capacity seemed to be less severely affected by eutrophication and high suspension loads.

Paleoenvironmental conditions marked by accelerated weathering and nutrient transfer have been reconstructed for the Early Cretaceous (e.g., Weissert, 1989; Föllmi et al., 1994). While eutrophication of Cretaceous oceans caused widespread carbonate platform growth crises and repeated carbonate platform drowning (Hallock and Schlager, 1986) the impact of eutrophication seemed to be less dramatic for Late Jurassic carbonate platforms. No major platform drowning comparable to the Cretaceous drowning events is reported for the Late Jurassic.

A monsoonal-controlled rainfall pattern with rainfall belts shifted to mid latitudes may explain why Late Jurassic carbonate platforms growing in low latitudes seemed to experience less severe growth crises than the Cretaceous platforms growing under increasingly zonal climate conditions. Nutrient poisoning or excessive shedding of siliciclastics are regarded as major factors resulting in the Cretaceous platform growth crises (e.g., Hallock and Schlager, 1986; Schlager, 1989; Föllmi et al., 1994). A comparison of the Late Jurassic rainfall pattern with the Cretaceous rainfall pattern suggests that the transition from a dominantely monsoonal climate to a more zonal climate in the Cretaceous had its impact on carbonate platform history. As a consequence, intensified weathering and rainfall may have had a stronger effect on low latitude carbonate platforms in the Cretaceous, while Late Jurassic carbonate platforms growing in arid low latitude belts (Hallam, 1992) were less severely affected by weathering debris. Thrombolite formation and replacement of coral reefs by sponge reefs may serve as indicators of most extreme environmental stress in Late Jurassic coastal regions.

The differing response mechanisms of the Late Jurassic and Early Cretaceous oceanic carbon pumps to climate change are reflected in the C org and C carb accumulation rate estimates given by Budyko et al. (1987). They conclude that the Late Jurassic was a time of high carbonate *and* high C org accumulation rates resulting in an accumulation ratio C carb:C org = 6:1 while the Early Cretaceous was marked by high C org and lowered C carb accumulation rates (C carb:C org = 4.7). Another expression of varying Late Jurassic–Early Cretaceous C carb burial rates is given by the amplitude of the C-isotope excursions in the Late Jurassic–Early Cretaceous C-isotope record (Fig. 5).

9. Conclusions: in the footsteps of Chamberlin and Salisbury

Paleoclimate simulations suggest that the Late Jurassic was marked by elevated atmospheric CO_2 levels (e.g. Moore et al., 1992). Conventional paleoclimate indicators such as the distribution of coals and evaporites can best be explained with an atmosphere enriched in greenhouse gases and with a monsoonal climate (Parrish, 1993). In this study we made an attempt to trace the history of two major elements of the biosphere, the water cycle and the carbon cycle, through the Late Jurassic. The carbon isotope data suggest that the mode of carbon cycling varied considerably through the

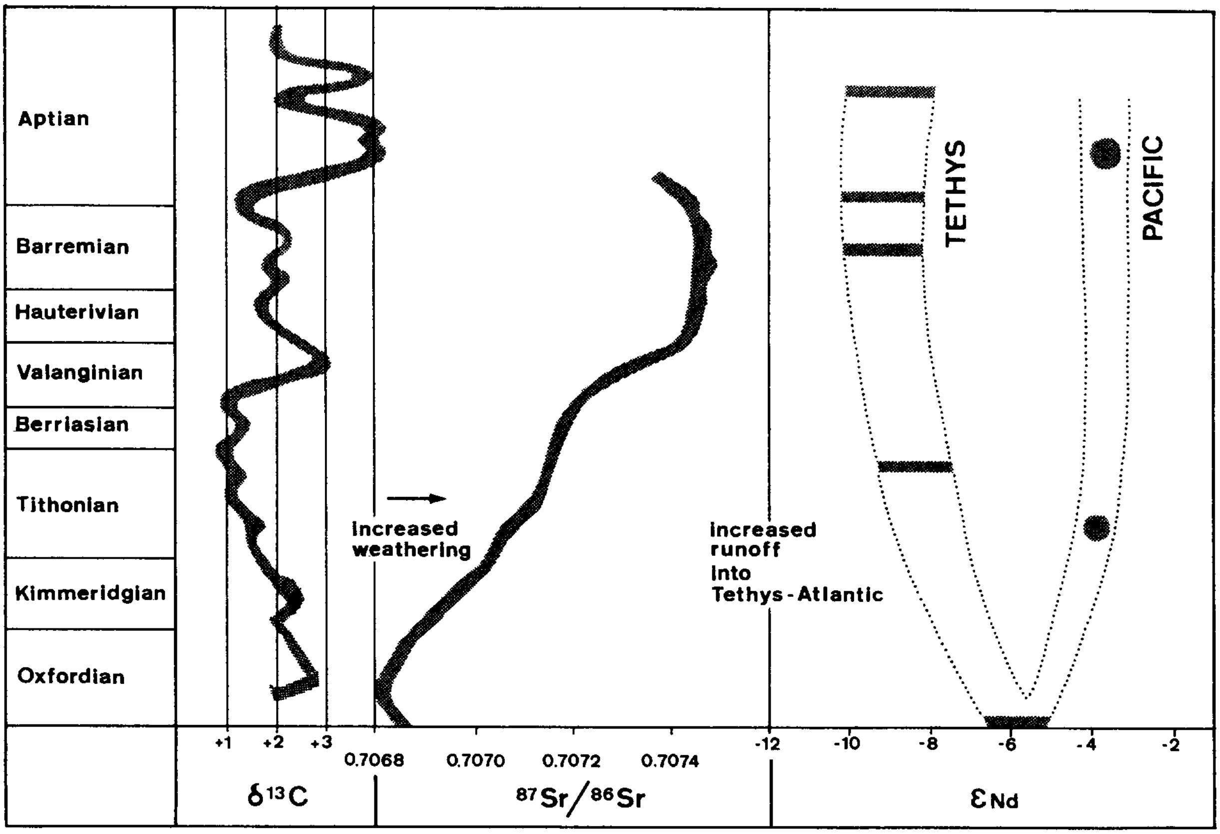

Fig. 7. Three tracers of paleoenvironmental change: The C-isotope curve records environmentally controlled fluctuations in Corg/Ccarb burial (Weissert and Lini, 1991; this study). The Sr-isotope record (Jones et al., 1994) is less sensitive to paleoenvironmental change than the C-isotope curve. It reflects a major climate reorganisation and an increase in global weathering and runoff between mid-Jurassic and mid-Cretaceous. The few available data on the εNd-stratigraphy (Stille et al., 1993) may reflect limited water exchange between the Atlantic and Pacific Oceans combined with increased runoff into the Tethys–Atlantic Ocean.

time window of interest. Conventional interpretation of the carbonate carbon isotope record suggests that the C org/C carb "export ratio" was increased during Late Jurassic with peaks falling into the Middle to Late Oxfordian and the early Late Kimmeridgian. Fluctuations in the Late Jurassic C org/C carb burial ratio were caused by (a) increased organic carbon accumulation rates and (b) changing efficiency of the oceanic carbonate carbon pump. The Middle to Late Oxfordian carbon isotope curve not only corresponds to an episode of elevated organic carbon accumulation rates but to times of growth crises of Northern Tethyan carbonate platforms reflected in widespread occurrence of siliceous sponge reefs with thrombolites. The Kimmeridgian–Tithonian C-isotope curve data only poorly reflect episodes of globally increased organic carbon accumulation. A high global sealevel and paleoceanographic conditions were favourable for carbonate platform growth at times of high organic carbon burial rates. Because of fluctuating carbonate carbon burial rates the Late Jurassic C-isotope record cannot be used as an accurate proxy for organic carbon accumulation rates.

If we try to integrate our information gained on the impact of water cycling, weathering and erosion on the Late Jurassic carbon cycling we may draw the following scenario (Fig. 7):

(1) Intensified Late Jurassic water cycling was

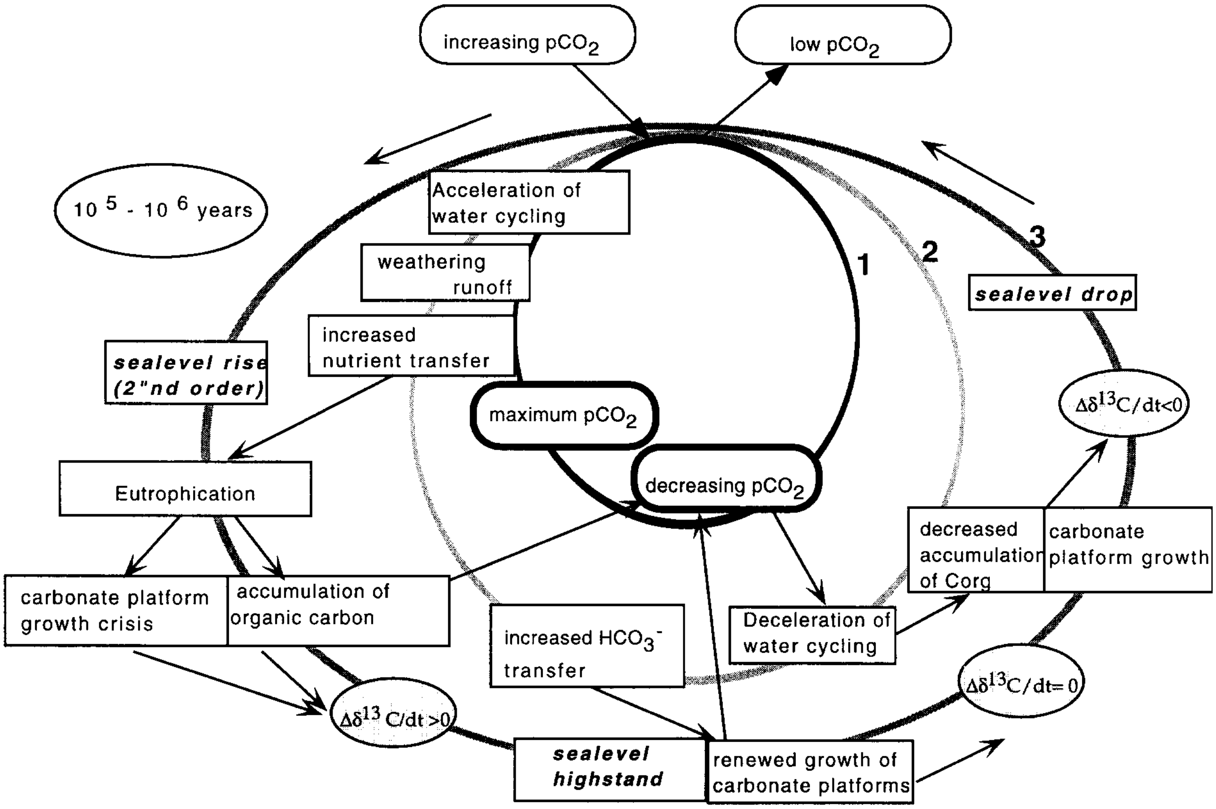

Fig. 8. Our model for the link between Late Jurassic global water cycle and global carbon cycle as it is inferred from available sedimentological and geochemical information.

triggered by elevated atmospheric carbon dioxide levels. The monsoonal climate controlled the Late Jurassic rainfall pattern. (2) Warm and humide climate favoured enhanced chemical weathering rates. Excessive carbon dioxide was transformed into alkalinity and nutrients were mobilized. (3) Increased chemical weathering and erosion rates resulted in an increased nutrient and alkalinity transfer from continents into oceans. The widespread occurrence of mature siliciclastics which were deposited in Late Jurassic oceans could serve as positive evidence for accelerated water cycling under monsoonal greenhouse conditions. (4) While the efficiency of the biological carbon pump was favoured by increased supply of nutrients the carbonate carbon pump was weakened in areas of eutrophication and elevated suspension levels. The Oxfordian C-isotope curve records the intensification of the biological carbon pump and the corresponding weakening of the carbonate carbon pump with a $\Delta\delta^{13}C/dt \gg 0$; (5) Elevated export of organic carbon and hence of excessive atmospheric carbon dioxide was favoured by biological carbon pumping, by increased sediment accumulation rates at times of enhanced weathering rates, and by slow renewal of water masses in such restricted ocean basins as the Jurassic North Sea. (6) Increased transfer of alkalinity from continents to oceans was ultimately compensated by elevated calcium carbonate sedimentation on inundated shelves. The gradual replacement of sponge reefs by coral reefs in the Kimmeridigan to Tithonian Northern Tethys may serve as a document of this change. The stabilisation of the Kimmeridgian C-isotope values at positive numbers records the growing efficiency of the carbonate carbon pump. (7) Through aggradation and progradation of carbonate platforms during the Late Jurassic sea level highstand excessive alkalinity and hence atmospheric CO_2 was stored in carbonates. A decreasing C org/C carb burial ratio is reflected in progressively lower Tithonian C-isotope values. (8) A lowering of atmospheric CO_2 levels due to accelerated chemical weathering and/or due to decreased volcanic and metamorphic degassing resulted in a deceleration of water cycling, weathering and nutrient cycling. If the change in the C-isotope curve at the end of the Jurassic coincided with a reorganisation of the global climate system and a transition to drier conditions (Weissert and Channell, 1989) was related to a collapse of the Jurassic monsoonal climate remains to be investigated. The remarkable differences in growth histories of Late Jurassic and Early Cretaceous carbonate platforms indicate that not only fluctuations in atmospheric carbon dioxide levels but a transition from a Jurassic mosoonal climate to a Cretaceous zonal climate had an impact on the mode of carbon cycling.

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