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An isotopic appraisal of the Late Jurassic greenhouse phase in the Russian Platform

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ABSTRACT

Oxygen- and carbon-isotope ratios have been determined from Late Jurassic (Callovian–Volgian) belemnites from three locations on the Russian Platform (Gorodischi, Khanskaya Gora and Marievka). All samples were examined by means of trace element geochemistry and petrography in order screen for diagenetic alteration. Oxygen and carbon isotopes from well-preserved belemnites range from −2.24 to −0.09‰ and −0.57 to 1.77‰ respectively. Oxygen isotopes, if interpreted in terms of temperature, reveal a rise of temperatures during the Oxfordian–Early Kimmeridgian and indicate a prolonged episode of warmth during the Kimmeridgian–Volgian. The isotope data only equivocally reflect a number of significant changes in Boreal–Tethyan ammonite assemblages. A positive carbon isotope excursion is observed within the Volgian, but not seen within composite carbon-isotope stratigraphies of the western Tethys. Hence the Jurassic may have been characterised by regional $\delta^{13}C$ excursions related to non-simultaneous organic matter deposition resulting from localised ponding, semi restricted ocean circulation and a lack of tidal mixing.

1. Introduction

Records of ocean temperatures in the Northern Hemisphere based upon the isotopic thermometry of fish and shark tooth enamel (Lécuyer et al., 2003; Dromart et al., 2003) indicate a severe cooling and subsequent rapid warming during the middle to Late Jurassic transition. For this reason Dromart et al. (2003) suggested that the middle to Late Jurassic transition may represent one of the major turning points of the climate history of the Earth. A number of compilations of Jurassic isotopic data (largely belemnite-derived) (e.g. Veizer et al., 1999; Barskov and Kiyashko, 2000; Jenkyns et al., 2002; Veizer, 2005) are supportive of this possible icehouse–greenhouse transition. Such isotopic databases frequently consist, however, of data from numerous dispersed locations where presumably potential differences exist with respect to temperature and the isotopic composition of seawater, hence making any global palaeotemperature reconstruction inherently complex. Certainly the Late Jurassic and in particular the Kimmeridgian has been identified as a period of time when temperatures reached a maximum (e.g. Frakes, 1979; Valdes and Sellwood, 1992; Abbink et al., 2001). Detailed isotopic records through this potential greenhouse interval are, however, limited (c.f. Price and Grocke, 2002; Gröcke et al., 2003; Wierzbowski, 2004; Zakharov et al., 2005). This study presents new (belemnite-derived) isotopic data from the Kimmeridgian–Volgian of the Russian Platform (Gorodischi, Khanskaya Gora and Marievka) combined with data from previous studies (also from the Russian Platform). A comprehensive ammonite zonation permits these data to be placed within a recognized and detailed biostratigraphical scheme.

2. Geological setting

During the Late Jurassic, the Russian Platform was located between palaeolatitudes ~35–50°N (Fig. 1; Smith et al., 1994; Thierry et al., 2000). Based on the palaeogeographic reconstructions of Sazonova and Sazanov (1967), land areas may have existed to the east and west of the study area, with marine connections to the Boreal and Tethyan seas. The width of the basin varied through time (Baraboshkin, 1997) but in the Late Jurassic was about 1200 km east to west and over 2000 km north to south.

The succession of Gorodischi village (25 km north of Ulyanovsk, Fig. 1) represents the stratotype of the Volgian (Gerasimov and Mikhailov, 1966) and ranges from the Kimmeridgian Eudoxus Zone to the Nodiger Zone in the Upper Volgian (Hantzpergue et al., 1998; Rogov, 2002; 2004, Fig. 2). Notably the base of Volgian and Tithonian are considered by some authors to be coincident (but see discussion by Scherzinger and Mitta, 2006). The succession is exposed over a distance of 15 km along the right bank of the Volga River and was first described by Murchison et al. (1845). Sediments of the lowermost ammonite zone seen (Eudoxus Zone) are composed of grey calcareous clays that locally grade into marl and yields a number of ammonites including *Aulacostephanus eudoxus*, *Sutneria* aff. *cyclodorsata*, *S. ex* gr. *Eumela*, *Aspidoceras quercynum*, *Discosphinctoides* sp., *Tolvericeras* cf. *sevogodense* and *Amoeboceras* spp. (Hantzpergue et al., 1998; Rogov, 2002). The overlying succession of the Autissiosorensis Zone (Fig. 2) is composed of a series of calcareous bioturbated light-grey clays locally

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Fig. 1. Palaeogeographic setting of the Russian Platform during the Tithonian (modified from Smith et al., 1994; Thierry et al., 2000). Inset shows the locations of Ulyanovsk, Syzran and Orenburg.

bearing calcareous concretions. These clays yield belemnites, bivalves and gastropods which frequently coexist with ammonites including *A. autissiosorensis* and its dimorphs *A.* cf. *kirghisensis*, *A.* ex gr. *undorae* (Hantzpergue et al., 1998) as well as numerous *Sarmatisphinctes*. The part of the succession assigned to the Klimovi Zone consists of alternating grey and dark grey bioturbated calcareous clays (Fig. 2). These clays have yielded belemnites, bivalves and gastropods and pyritized ammonites (Rogov, 2002). The overlying ~1 m of sediments contains the ammonites *Ilowaiskya* cf. *sokolovi*, and *Subdichotomoceras* cf. *subcrassum* characteristic of the Sokolovi Zone. The Pseudoscythica Zone contains the '*neoburgensis* horizon' (of Rogov, 2002, 2005), which is rich in small *Hibolithes* belemnites and is dominated by Submediterranean ammonites such as *Anaspidoceras neoburgensis*. The sediments assigned to the *puschi* horizon and Panderi Zone are grey calcareous bioturbated clays dominated by Subboreal ammonites belonging to the *Pseudovirgatites–Zaraiskites* lineage with also *Buchia* sp., *Liostrea* sp., *Astarte* sp., *Dicraloma* sp. and *Oxytoma* sp. The upper part of the succession, comprising calcareous mudstones interbedded with a number of prominent black shale horizons, is overlain by a condensed (~2 m) silty–sand unit with abundant phosphatic nodules (Fig. 2). Invertebrates from this highly fossiliferous part of the succession include belemnites, ammonites and bivalves (Kuleva et al., 1996; Hantzpergue et al., 1998). The black shales are extremely enriched in organic carbon and reach a maximum of 40–50 wt.% (Hantzpergue et al., 1998; Riboulleau et al., 2003). Equivalent organic-rich deposits to those seen at Gorodischi are widely distributed on the eastern and middle parts of the Russian Platform, with a total outcropping and sub cropping area of more than 100 000 km^2 (Vishnevskaya et al., 1999; Riboulleau et al., 2003; Gavrilov et al., 2008). The black shales of Gorodischi are also the lateral equivalent of the lower part of the Bazhenov Formation, the main source rock of the west Siberian oil fields (Riboulleau et al., 2003).

A further 13 belemnite samples were obtained from Belyaevka Village, Khanskaya Gora located ~60 km southeastwards from Orenburg (Fig. 1), on the left bank of Berdjanka river. They are derived from the Upper Kimmeridgian Eudoxus Zone to the Lower Volgian Tenuicostatum Zone. This part of the succession is ~9 m thick and consists of interbedded silts and sandstones with chert concretions and yielded abundant ammonites described by Ilovaiskii and Florenskii (1941) and Mikhailov (1966). A number of belemnites (*Cylindroteuthis* sp.) were also obtained from this section (Table 1). A further 5 belemnites were obtained from the Marievka Village section, located 40 km west of Syzran (Fig. 1). This small section consisted of a 3 m thick unit of calcareous sands overlain by a phosphatic nodule-rich layer and ranged from the Nikitini to Nodiger Zone. Belemnites (*Acroteuthis* (*Microbelus*) sp.) were sampled from the Nodiger Zone only.

3. Materials and methods

Oxygen and carbon isotopic compositions have been determined from well preserved specimens of the belemnite genera *Cylindroteuthis*, *Pachyteuthis* (*P.*), *A.* (*Microbelus*), and *Hibolithes*. Where possible, multiple samples were collected from the same stratigraphic horizon. Diagenetic

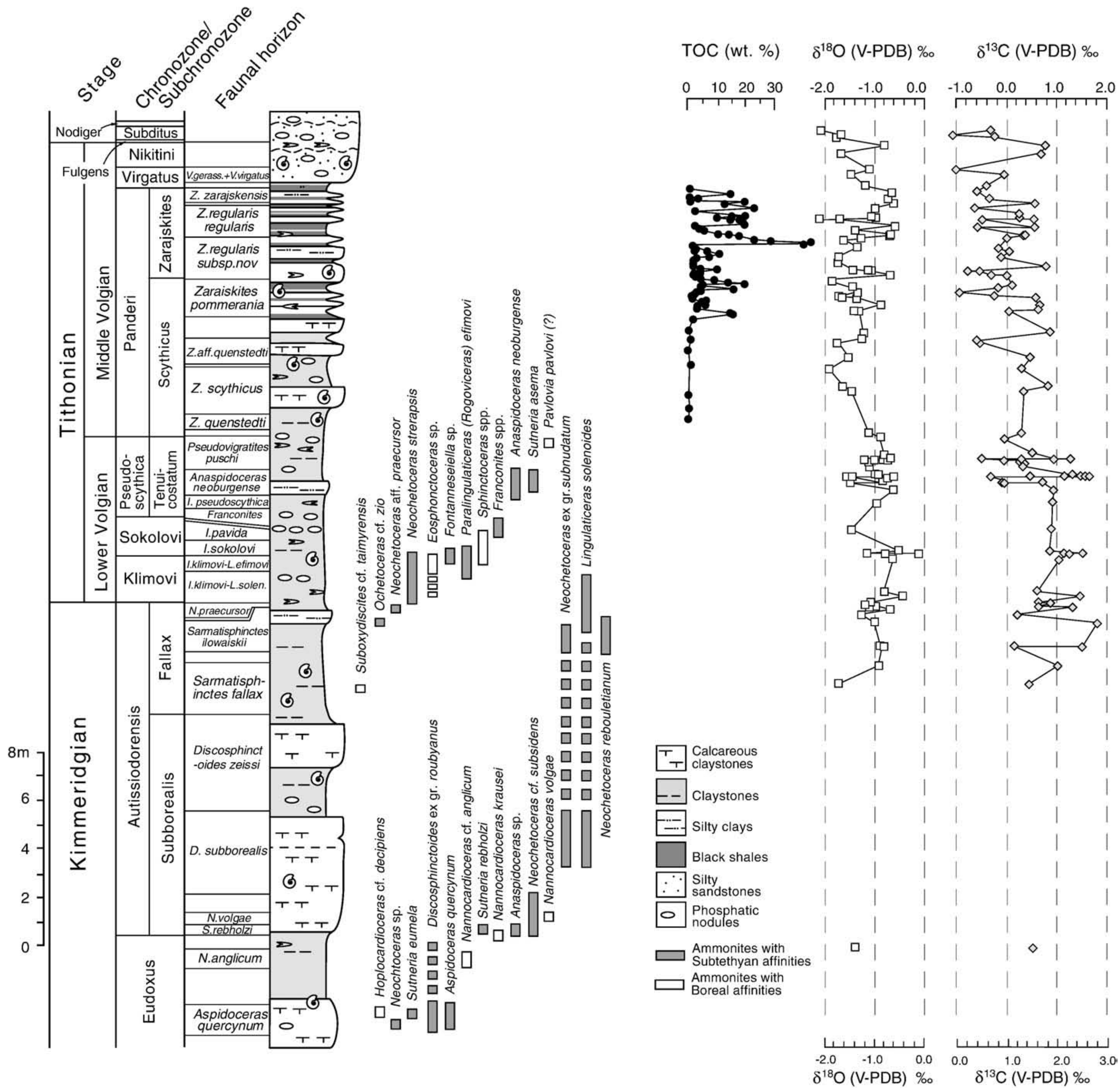

Fig. 2. Isotope variation during the Late Jurassic–earliest Cretaceous interval at Gorodischi (with additional data from Riboulleau et al., 1998; Gröcke et al., 2003). Generalised sedimentary log and biostratigraphy from field observations, Hantzpergue et al. (1998) and Rogov and Kiselev (2007). Total organic carbon from Riboulleau et al. (2003). Faunal horizons and ranges of ammonites with Boreal or Subtethyan affinities from Rogov et al. (2006) and Rogov and Kiselev (2007).

alteration of each of the belemnite samples was characterised by a combination of petrographic investigations and trace element geochemistry. Carbonate staining (Dickson, 1966) revealed partial replacement by ferroan calcite preferentially along the outermost concentric growth bands and associated with the apical line. Areas such as these were either removed prior to or avoided during subsampling. The remains were fragmented (sub-mm), washed in ultrapure water, and dried in a clean environment for subsequent isotopic analysis. Because of the possibility of incorporation of Mn and Fe from pore waters during diagenesis, all samples were analysed additionally for trace element contents. Subsamples for chemical analysis (Fe, Mn) were dissolved in concentrated nitric acid and analysed using a Perkin Elmer 3100 Atomic Absorption Spectrometer. Based upon analysis of duplicate samples reproducibility was better than ±3% of the measured concentration of each element. Carbon and oxygen isotopes were determined on a VG Instruments Optima Isotope Ratio Mass Spectrometer with a Multiprep Automated Carbonate System (at the University of Plymouth) using 200 to 300 micrograms of carbonate. The $\delta^{13}C$ and $\delta^{18}O$ values obtained were calibrated against NBS-19. Analytical reproducibility of the measurements is ±0.2‰ based upon replicate analyses. The $\delta^{18}O$ and $\delta^{13}C$ data are reported in the conventional delta notation with respect to V-PDB. In determining calcite palaeotemperatures, the equation of Epstein et al. (1953) as modified by Anderson and Arthur (1983) was used, as it was based primarily on molluscan isotope data. This equation

Table 1
Isotopic and elemental compositions of belemnite genera analysed from Gorodischi, Khanskaya Gora and Marievka. Indet = indeterminable

Sample number	Location	Ammonite zone	Genus	$\delta^{13}C$	$\delta^{18}O$	Fe (ppm)	Mn (ppm)
Gor16	Gorodischi	*A. autissiodorensis*	*Cylindroteuthis* sp.	0.39	−1.72	24	9
Gor21	Gorodischi	*A. autissiodorensis*	*Cylindroteuthis* sp.	0.96	−0.94	14	14
Gor8	Gorodischi	*A. autissiodorensis*	*Cylindroteuthis* sp.	0.08	−0.86	8	7
Gor13	Gorodischi	*A. autissiodorensis*	*Cylindroteuthis* sp.	1.46	−0.88	8	5
Gor3	Gorodischi	*A. autissiodorensis*	*Cylindroteuthis* sp.	1.77	−1.02	8	13
Gor26	Gorodischi	*A. autissiodorensis*	*Cylindroteuthis* sp.	0.15	−1.29	73	24
Gor27*	Gorodischi	*A. autissiodorensis*	*Hibolithes* sp.	2.23	−0.70	272	10
Gor24	Gorodischi	*A. autissiodorensis*	*Cylindroteuthis* sp.	1.25	−0.97	15	82
Gor25	Gorodischi	*A. autissiodorensis*	*Cylindroteuthis* sp.	0.56	−1.16	14	94
Gor11	Gorodischi	*A. autissiodorensis*	*Cylindroteuthis* sp.	0.85	−1.20	26	36
Gor7	Gorodischi	*A. autissiodorensis*	*Cylindroteuthis* sp.	0.58	−1.08	7	20
Gor28	Gorodischi	*A. autissiodorensis*	*Pachyteuthis* sp.	1.44	−0.43	3	2
Gor17	Gorodischi	*A. autissiodorensis*	*Cylindroteuthis* sp.	0.55	−0.82	22	42
Gori 6d2	Gorodischi	*I. klimovi*	*Cylindroteuthis* sp.	1.00	−0.63	26	8
Gori 6D10a	Gorodischi	*I. klimovi*	*Cylindroteuthis* sp.	1.47	−1.15	26	6
Gori 6d10b	Gorodischi	*I. klimovi*	*Pachyteuthis* (*P.*) sp.	1.08	−0.09	15	5
Gori 6d10c	Gorodischi	*I. klimovi*	*Pachyteuthis* (*P.*) sp.	1.19	−0.81	18	6
Gori 6D35	Gorodischi	*I. klimovi*	*Cylindroteuthis* sp.	0.80	−0.52	19	7
Gori 7D10	Gorodischi	*I. sokolivi*	*Cylindroteuthis* sp.	0.84	−1.48	16	6
Gori 8D5	Gorodischi	*I. pseudoscythica*	*Cylindroteuthis* sp.	0.86	−0.99	24	10
Gor14	Gorodischi	*I. pseudoscythica*	*Hibolithes* sp.	0.89	−0.61	44	3
Gor12	Gorodischi	*I. pseudoscythica*	*Hibolithes* sp.	−0.18	−0.87	21	35
Gor2	Gorodischi	*I. pseudoscythica*	*Hibolithes* sp.	0.66	−1.52	22	25
Gor20	Gorodischi	*I. pseudoscythica*	*Pachyteuthis* (*P.*) sp.	−0.44	−0.76	4	5
Gor4	Gorodischi	*I. pseudoscythica*	*Pachyteuthis* sp.	1.25	−0.97	36	2
Gor18	Gorodischi	*I. pseudoscythica*	*Cylindroteuthis* sp.	0.22	−1.11	72	2
Gor5	Gorodischi	*I. pseudoscythica*	*Cylindroteuthis* sp.	0.31	−1.09	8	2
Gor6	Gorodischi	*I. pseudoscythica*	*Pachyteuthis* sp.	0.13	−1.09	57	4
Gor30	Gorodischi	*I. pseudoscythica*	*Cylindroteuthis* sp.	−0.11	−0.98	3	2
Gor1	Gorodischi	*I. pseudoscythica*	*Pachyteuthis* sp.	0.20	−1.24	5	3
Gor22	Gorodischi	*I. pseudoscythica*	*Cylindroteuthis* sp.	1.26	−0.82	16	2
Gor19	Gorodischi	*I. pseudoscythica*	*Cylindroteuthis* sp.	0.91	−0.81	8	2
Gor10	Gorodischi	*D. panderi*	*Cylindroteuthis* sp.	0.45	−0.82	4	19
Gor31*	Gorodischi	*V. virgatites*	*Hibolithes* sp.	1.62	−5.27	2777	48
Gor9*	Gorodischi	*V. virgatites*	*Hibolithes* sp.	1.53	−5.36	154	31
3/1u65	Gorodischi	*D. panderi*	*Cylindroteuthis* sp.	−0.56	−1.28	16	5
3/1d200	Gorodischi	*D. panderi*	*Cylindroteuthis* sp.	0.71	−2.29	14	6
2/1u390	Gorodischi	*D. panderi*	*Cylindroteuthis* sp.	0.58	−0.66	13	5
2/1u250	Gorodischi	*D. panderi*	*Cylindroteuthis* sp.	−0.63	−1.02	21	8
2/1u240	Gorodischi	*D. panderi*	*Cylindroteuthis* sp.	−0.10	−1.03	17	6
2/1u82	Gorodischi	*D. panderi*	*Cylindroteuthis* sp.	0.72	−1.59	17	5
1/2d60	Gorodischi	*D. panderi*	*Cylindroteuthis* sp.	−0.76	−1.23	19	8
1/8a	Gorodischi	*I. pseudoscythica*	*Cylindroteuthis* sp.	1.77	−1.11	40	7
1/8b	Gorodischi	*I. pseudoscythica*	*Cylindroteuthis* sp.	1.50	−1.64	28	36
1/8c	Gorodischi	*I. pseudoscythica*	*Cylindroteuthis* sp.	0.52	−1.61	24	4
1/8d	Gorodischi	*I. pseudoscythica*	*Cylindroteuthis* sp.	1.46	−0.63	35	5
B/25-5	Khanskaya	*tenuicostatum*	*Cylindroteuthis* sp.	0.63	−1.45	27	7
B/16-10	Khanskaya	*I. pseudoscythica*	Indet.	0.44	−1.30	41	7
B/15-5*	Khanskaya	*I. pseudoscythica*	*Cylindroteuthis* sp.	−0.26	−2.60	1565	16
B/11-10	Khanskaya	*I. sokolivi*	*Cylindroteuthis* sp.	1.03	−1.62	25	6
B/4-5	Khanskaya	*I. klimovi*	Indet.	0.73	−1.42	112	20
B/3-10	Khanskaya	*I. klimovi*	*Cylindroteuthis* sp.	1.20	−2.24	15	2
B/2-17	Khanskaya	*Autissiodorensis*	*Cylindroteuthis* sp.	−0.08	−2.27	110	20
B/2-10	Khanskaya	*Autissiodorensis*	*Cylindroteuthis* sp.	−0.29	−1.46	20	4
B1-10	Khanskaya	*Autissiodorensis*	?*Cylindroteuthis* sp.	0.62	−1.77	28	4
A/7	Khanskaya	*Autissiodorensis*	*Cylindroteuthis* sp.	1.22	−1.28	40	4
A/6	Khanskaya	*Autissiodorensis*	*Cylindroteuthis* sp.	1.23	−2.03	39	10
A/5	Khanskaya	*eudoxus*	*Cylindroteuthis* sp.	0.53	−1.99	17	4
A/4	Khanskaya	*eudoxus*	Indet.	0.81	−1.79	86	25
M-1*	Marievka	*C. nodiger*	*A.* (*Microbelus*) sp.	−0.31	−1.28	187	9
M-2	Marievka	*C. nodiger*	*A.* (*Microbelus*) sp.	−0.57	−2.09	97	33
M-3	Marievka	*C. nodiger*	?*Acroteuthis* sp.	0.49	−1.96	38	6
M-4	Marievka	*C. nodiger*	*A.* (*Microbelus*) sp.	0.15	−1.77	7	5
M-5*	Marievka	*C. nodiger*	*A.* (*Microbelus*) sp.	−0.33	−2.09	252	12

* Deemed diagenetically altered and not analysed further.

expresses the oxygen isotopic composition of the water (δ_{water}), directly relative to the standard mean ocean water (SMOW) standard:

$$T(°C) = 16.0 - 4.14(\delta_{calcite} - \delta_{water}) + 0.13(\delta_{calcite} - \delta_{water})^2$$

where $\delta_{calcite}$ equals the oxygen isotopic composition of the calcite with respect to the VPDB international standard and δ_{water} equals the oxygen isotopic composition of the water ($\delta_{seawater}$) from which the calcite was precipitated with respect to the SMOW standard. A

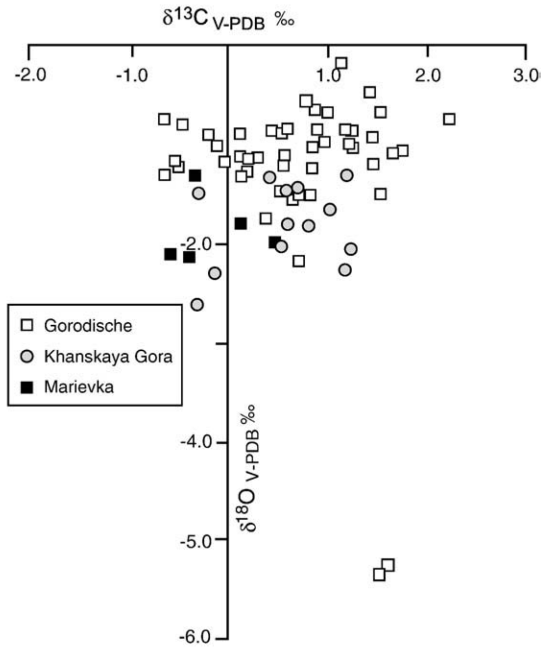

Fig. 3. Cross-plot of $\delta^{18}O$ and $\delta^{13}C$ values of belemnite samples from Gorodischi, Khanskaya Gora and Marievka.

$\delta_{seawater}$ of −1.0‰ (SMOW) is thought to be appropriate for assumed non-glacial periods (Shackleton and Kennett, 1975).

4. Results

Trace element and isotope data for the belemnites are presented in Table 1 and Figs. 2, 3. This dataset complements those already obtained previously from diagenetically screened belemnites from Gorodischi (Riboulleau et al., 1998; Gröcke et al., 2003). The belemnites sampled in this study were mostly translucent and retained the primary concentric banding that characterise belemnite rostra. Relatively low Mn concentrations are typically observed in modern calcitic molluscs (e.g. Milliman, 1974) and hence such values can be assumed to reflect well-preserved fossil material (cf. Riboulleau et al., 1998; Podlaha et al., 1998; Price and Mutterlose, 2004; Wierzbowski, 2004; McArthur et al., 2004; Price and Page, 2008). The typically low Mn (<100 ppm) and Fe (<150 ppm) values, recorded for most of the belemnites (Table 1), in conjunction with the petrographic evidence, are thus consistent with minimal diagenetic alteration. The higher amounts of Mn and Fe and occasional outliers displaying more negative $\delta^{18}O$ and $\delta^{13}C$ values noted in some of the belemnites (Table 1, Fig. 3) are regarded as an artefact of diagenetic alteration. Those samples where Fe concentrations were >150 ppm and Mn concentrations >100 ppm were excluded from further analysis. The belemnite data from Gorodischi (Fig. 2) displays a comparatively small range of $\delta^{18}O$ values (−0.09 to −2.11‰). The most positive values are seen within the Klimovi Zone and at the top of the Panderi Zone. $\delta^{13}C$ values show a larger variation (−1.20 to 1.77‰) and show an overall trend towards negative values from the Kimmeridgian through into the Volgian.

Within Fig. 4 the data from Gorodischi (this study; Riboulleau et al., 1998; Gröcke et al., 2003), Marievka and Khanskaya Gora (this study) have been combined with data from other locations in the Russian Platform, Makariev village (Riboulleau et al., 1998); Yaroslavl region and Voskresensk (Podlaha et al., 1998) and plotted using the Gradstein et al. (2004) timescale. Those data included from Riboulleau et al. (1998) and Podlaha et al. (1998) were also diagenetically screened. Yaroslavl is located 250 km north east of Moscow, Voskresensk is located 80 km south east of Moscow, whilst Makariev village is located 300 km northeast of Moscow (Fig. 1). The most positive $\delta^{18}O$ values are observed in the Upper Callovian and decrease to more negative values in the Oxfordian and Kimmeridgian. Through into the Volgian and Lower Cretaceous, the $\delta^{18}O$ values reveal surprisingly very little variation with the most negative values seen in the Middle and Upper Volgian. The most positive $\delta^{13}C$ values are also observed in the Upper Callovian–lowermost Oxfordian. A second peak is possibly observed in the Kimmeridgian (although based on only a few data points), followed by a gradual decrease of values into the Volgian. The trend is interrupted by a marked increase in values, seen in the Middle Volgian, notably in both data from Gorodischi and from Voskresensk.

5. Discussion

Increasingly negative $\delta^{18}O$ values in carbonates can be related to higher temperatures in environmental settings where continental ice volume over time is constant (and therefore invariable $\delta_{seawater}$) and evaporation or freshwater input are minor factors. Belemnites are also thought to secrete their carbonate in isotopic equilibrium with ambient seawater (e.g. Lowenstam and Epstein, 1954; Price and Mutterlose, 2004). This assumption is supported by the fact that $\delta^{18}O$ values of modern *Sepia* shells (a close relative of belemnites) also appear close to equilibrium fractionation (Bettencourt and Guerra, 1999; Rexfort and Mutterlose, 2006). Although $\delta^{13}C$ signatures from *Sepia* shells are likely to be affected by vital effects (Rexfort and Mutterlose, 2006), useful information can still be gained from their carbon isotopic analysis. Hence assuming oxygen isotope equilibrium, a rise in temperature is observed from the Callovian into the Oxfordian from ~4 to 18 °C (Fig. 4) (see also Hoffman et al., 1991; Barskov and Kiyashko, 2000; Malchus and Steuber, 2002). Dromart et al. (2003) have identified this rapid rise in temperatures as evidence of a glacial to interglacial transition. This trend towards more negative $\delta^{18}O$ values and rise in temperature is seen to continue until the Oxfordian–Kimmeridgian boundary (although the data for the Middle and Upper Oxfordian are a little sparse). A similar trend is observed by Wierzbowski (2004) obtained from belemnite-derived isotope measurements from central Poland and southern Germany. Within the database of Veizer et al. (1999) a larger change in oxygen isotope values is observed through the Callovian to Early Kimmeridgian and hence a greater magnitude change in temperature is possibly implied. Within this latter study, isotope data was derived from numerous locations (in both the northern and southern hemispheres) where presumably potential differences exist with respect environmental parameters and latitudinal variations in temperature making any global palaeotemperature reconstruction inherently complex.

The isotope temperature values for the Callovian–Oxfordian (derived from the data of Podlaha et al., 1998; Riboulleau et al., 1998) are worthy of closer scrutiny. The $\delta^{18}O$ values of belemnites imply temperatures as low as 4 °C (Fig. 4) or a ~2–3 °C warmer if a $\delta_{seawater}$ values of −0.5 to 0.0‰ are used (i.e. adjusted to the presence of some high latitude ice, Dromart et al., 2003). These data, derived from the middle part of the Russian Platform (Fig. 1), would therefore suggest locally a very frigid climate and ice at the poles. Given that changes in $\delta^{18}O_{seawater}$ can be related to salinity, such low temperatures could be an artefact of significant fluctuations in water mass chemistry i.e. more saline conditions or related to depth preferences of belemnites. Increases in salinity, for example to ~38‰ would increase temperatures by ~5 °C, although such high salinities in conjunction with relatively cool temperatures are likely to be incompatible. Alternatively if belemnites had been deep-ocean dwellers, cooler temperatures may reflect these deeper depths. This scenario appears less likely because of the known (relatively shallow) depth of the Russian Platform and also as most authors (e.g. Westermann, 1973; Price and Mutterlose, 2004; Wierzbowski, 2004) would advocate that most belemnites tended to inhabit the shallower shelf environments.

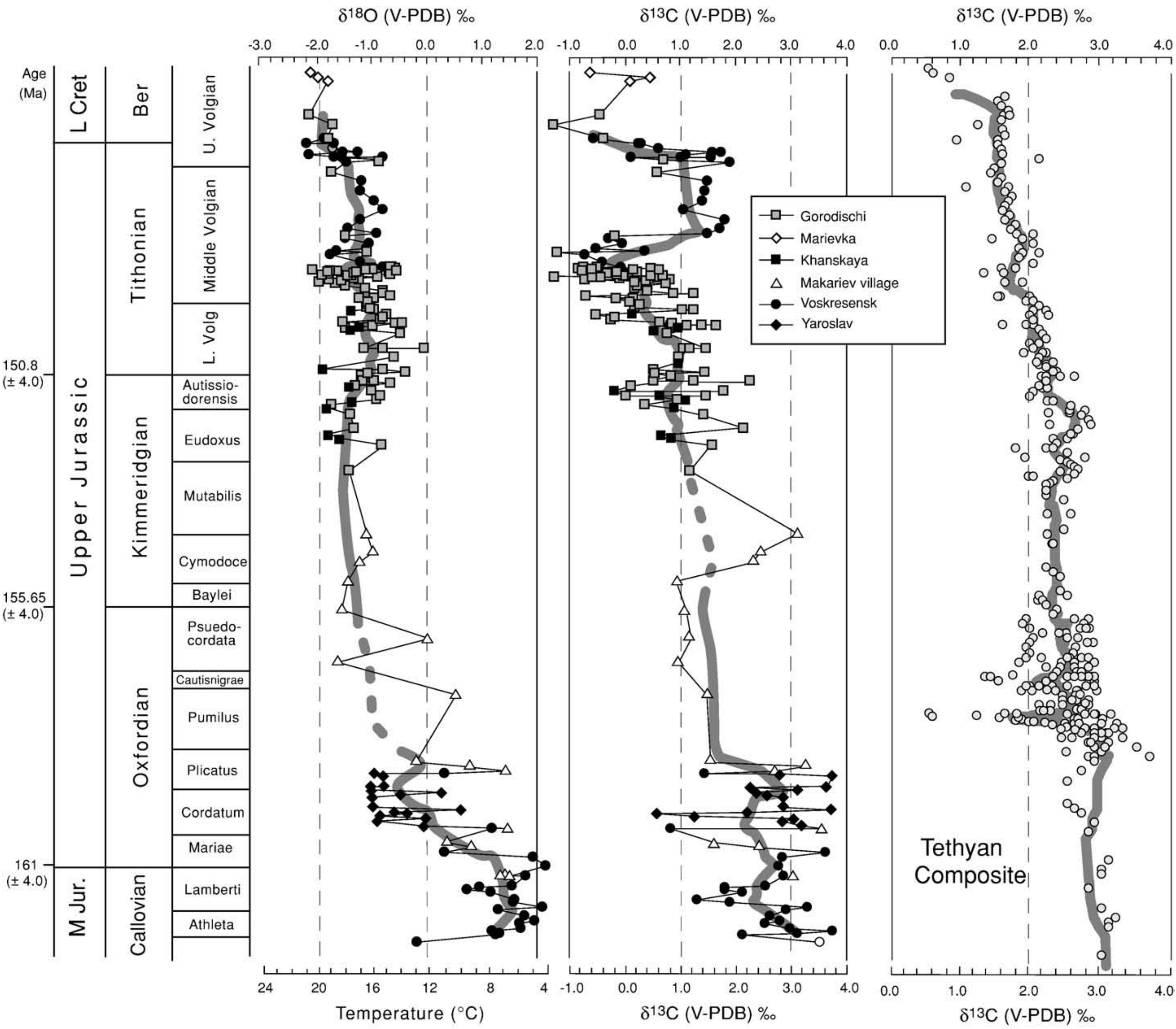

Fig. 4. Summary of belemnite isotope data from the Russian Platform and 10-point moving average (data from this study; Yaroslavl, Voskresensk: Podlaha et al., 1998; Makariev village: Riboulleau et al., 1998). Ages from Gradstein et al. (2004) and Late Jurassic composite Tethyan δ^{13}C curve from Jenkyns (1996); Dromart et al. (2003); Padden et al. (2002); Weissert and Channell (1989) and Weissert and Mohr (1996).

The data presented for the Kimmeridgian–Volgian show surprisingly little variation (and considerably less scatter than some compilations of similar data, e.g. Veizer et al., 1999; Jenkyns et al., 2002) with δ^{18}O values becoming steadily more negative. This is despite the distance between sites (up to 1600 km apart, Fig. 1) and potentially different oceanographic settings. The Late Jurassic greenhouse and during the Kimmeridgian–Tithonian in particular, have been identified as a period of time when temperatures reached their maximum development (e.g. Frakes, 1979; Valdes and Sellwood, 1992; Price, 1999; Abbink et al., 2001). What is notable within this longer-term trend is a lack of a clear peak of temperatures. Instead the oxygen isotope data indicate a prolonged episode of gradual warming. Zakharov et al. (2005) also present belemnite-derived stable isotope data from the Kimmeridgian of the Subpolar Urals. What is noteworthy is that a very similar oxygen isotopic range (−1.25 to −2.50‰) is documented. Zakharov et al. (2005) notes the high palaeotemperatures of their study and in part accounts for them as being influenced by freshwater runoff (thereby artificially inflating apparent temperatures). Nevertheless, given that a similar range of oxygen isotope values (and hence temperatures) are observed within the Russian Platform, this may suggest that more than a local effect i.e. the observed trends may reflect actual warmth.

Jurassic carbon-isotope stratigraphies of the Western Tethys (e.g. Weissert and Channell, 1989; Jenkyns, 1996; Bartolini et al., 1996; Weissert and Mohr, 1996; Padden et al., 2002; Dromart et al., 2003) record positive anomalies during the Callovian, middle Oxfordian and Kimmeridgian and a decrease towards the Jurassic–Cretaceous boundary (Fig. 4). The most positive carbon isotope values in the Kimmeridgian coincide with extensive organic deposition during the Eudoxus–Hudlestoni Zones (Morgans-Bell et al., 2001). Superimposed upon these trends are possibly two negative excursions within the middle Oxfordian (Transversarium Zone) that have been explained by paleoceanographic changes, significant volcanic activity or a sudden release of methane from buried gas hydrate (Padden et al., 2001).

Positive (albeit highly variable values) are observed in the belemnite data for the Callovian and Oxfordian. A decline in $\delta^{13}C$ values seen through the Kimmeridgian and Lower Volgian is interrupted by a marked positive excursion in the Upper Volgian before a return to pre-excursion values across the Jurassic–Cretaceous boundary, notably seen in the data from both this study and Podlaha et al. (1998). Carbon-isotope excursions have been linked directly with episodes of increased burial of organic carbon attributed to enhanced preservation under reduced O_2 conditions (e.g. Bralower and Thierstein, 1984) or driven by changes in surface water productivity (delivering more organic carbon to the sea floor, e.g. Weissert, 1989; Kessels et al., 2003). Models have been presented whereby the leaching of nutrients on coastal lowlands during a rise in sea level possibly triggered by globally warmer and more humid climates, resulted in increased ocean fertilisation, productivity and an expansion of the oxygen minimum zone (e.g. Erbacher and Thurow, 1997, Röhl et al., 2001). This in turn leads to both organic matter preservation and marine waters enriched in ^{13}C. As demonstrated here, the organic rich units of Gorodischi do not fall within the main positive segment of the $\delta^{13}C$ curve (Fig. 2). Such an observation may simply reflect the fact that the $\delta^{13}C$ record is a global signal whilst the black shale deposition is a local phenomenon. However, the Tethyan $\delta^{13}C$ record (Fig. 4), shows no major positive excursion in the Tithonian, meaning that a global excursion is questionable. In contrast a $\delta^{13}C$ decline is observed which is considered as evidence of increasingly oligotrophic conditions in the Tethyan seaway (e.g. Weissert and Channell, 1989). Accordingly the excursion may not be simply driven by increased carbon burial (cf. Menegatti et al., 1998). Instead preferential extraction of ^{12}C into the sedimentary carbon sink could be balanced by a flux of ^{12}C to the surface waters (e.g. upwelling of intermediate water or an intensified flux of riverine DIC) in order to provide steady state conditions for deposition of the organic rich levels, without a marked positive carbon isotope excursion. Likewise Weissert and Mohr (1996) also note that episodes of enhanced organic carbon burial during the Late Jurassic were not reflected by prominent positive excursions in the carbon isotope record. Weissert and Mohr (1996) suggest that carbon release from weathering must have more or less equalled carbon burial during this time and hence steady $\delta^{13}C$ values were maintained. Intensified riverine DIC input does not, however, appear a likely source of ^{12}C enriched waters given a semi-arid climate of the Late Jurassic Russian Platform (Ruffell et al., 2002).

Clay mineral abundances within the Gorodischi succession indicate that deposition of the (productivity driven) organic-rich units occurred during periods of increasing aridity associated with enhanced transport of atmospheric dust supplying iron (Ruffell et al., 2002; Riboulleau et al., 2003). Whilst, a pronounced $\delta^{13}C$ excursion is seen in the Middle Volgian, the $\delta^{18}O$ data show no discernable inflection in the curve suggestive of a marked change in temperature or water mass characteristics. If the carbon excursion and black shale deposition are linked, it may therefore be the case that although climate change (e.g. aridity) was an important productivity control, non climatic factors, such as preservation of organic matter may also be important (c.f. Kessels et al., 2003). These might include the semi enclosed nature of the basin leading to a lack of tidal mixing and stratification.

Studies of changes in ammonite assemblages in space and time in terms of faunal horizons have shown that in areas such as the Russian Platform, there have been significant oscillations from horizon to horizon that are easily traced regionally (Rogov et al., 2006; Rogov and Kiselev, 2007). Certain oscillations (for example, the abrupt replacement of a Subboreal–Submediterranean ammonite-dominant assemblages by a Boreal ammonite-dominant one at the base of Autissiodorensis Zone) can be traced from the Russian Platform to Poland and to the UK and thus could reflect global or subglobal events. The controls upon the changes seen in the ammonite assemblages may be ocean temperature variation, salinity or changes in palaeogeography (e.g. the presence or absence of seaways) (e.g. Rawson, 1973; Hallam, 1975; Price et al., 2000). Recently Rogov and Kiselev (2003) demonstrated that the positive shift in oxygen isotopes, interpreted as sharp drop in temperatures at the Callovian–Oxfordian boundary (e.g. Barskov and Kiyashko, 2000; Dromart et al., 2003, Fig. 4) was accompanied by an increase in the amount of Boreal ammonites in assemblages of the Russian Platform, the same trend also known from Western Europe (i.e. the 'Boreal Spread' of Arkell, 1956).

Changes in ammonite assemblages can be similarly compared with the stable isotope record from the Gorodischi section. At the Eudoxus–Autissiodorensis boundary a change from Subtethyan dominated ammonite assemblages to Boreal-dominated ones (e.g. Subtethyan *S. aff rebholzi* to Boreal *N. volgae* horizons). These cycles can be traced through Polish and English sections. Unfortunately we do not have sufficient belemnite data from this interval (there is an only one isotope determination made by Riboulleau et al. (1998) from a belemnite obtained from the upper part of the Eudoxus Zone seen in Fig. 2). Subtethyan faunas are also prevalent during the *neoburgense* faunal horizon (Rogov, 2004, 2005). Although repeated oscillations in stable isotopes derived from these subzones may reflect changes in water temperature, unambiguous shifts to more negative (warmer) oxygen isotope values coincident with Subtethyan faunas or vice-versa are not apparent (Fig. 2). Likewise, during the Panderi Zone numerous isotope oscillations showing higher palaeotemperatures in comparison with the Kimmeridgian–Volgian transition are again not reflected in structure of ammonite communities. Subtethyan derivates seem to be extremely rare, but instead the base of the Panderi Zone is marked by incursion of Boreal *Dorsoplanites* and *Pavlovia*, which make up the bulk of the ammonite assemblage in some sections (Khanskaya Gora, for example). Changes in ammonites faunas thus when compared to the isotope record appear to be more noticeable than any isotope trend.

The lack of a correlation between Boreal/Subtethyan ammonite incursions and oxygen isotope data interpreted in terms of temperature may suggest that either temperature is not a straightforward control upon migration or other factors are more significant in controlling migration routes and destinations. Similar scenarios have been described from the UK and Poland whereby oxygen isotope data do not correlate with the appearance of short pulses of warm or cold water cephalopod faunas (e.g. Price et al., 2000; Wierzbowski, 2002). Certainly warming and cooling episodes will not uniformly affect the surface of the globe. An increase in temperature in Tethys resulting in a displacement of faunas northwards into the Russian Platform would need to be accomplished with no marked temperature increase in the Russian Platform region itself. Typically a different pattern is observed, whereby when a temperature change occurs, warming is greatest in higher rather than lower latitudes. In terms of the Panderi Zone it may be speculated that perhaps major migrational pathways were blocked (e.g. the Brest strait was closed at this time and connection with the Caucasian Basin was restricted since ~Early–Middle Volgian times). Faunal exchange with the Caucasus basin may have been reduced after the expansion of the evaporitic belt in the Peri-Caspian area (e.g. Sazonova and Sazanov, 1967; Baraboshkin, 1999). The potential absence of the ammonite migrational pathways from Boreal basins to Tethys during the bulk of the Middle Volgian and Late Volgian leads to the well-known difficulties with Boreal–Tethyan correlation.

The gradual temperature increase (Figs. 2, 4) during the Middle–Upper Volgian is also poorly reflected in ammonite records. From the Nikitini Zone onwards numerous Arctic ammonites invaded the Middle Russian Sea (e.g. Baraboshkin, 1999; Kiselev and Rogov, 2005). Hence there is again the juxtaposition of Boreal elements, Arctic ammonites and warm palaeotemperature signals. The gradual shallowing of the Middle–Late Volgian sea may have led to a shift in the natural habitat of belemnitids to more shallow and warm surface water (see also Zakharov et al., 2005). Alternatively, a major limiting factor in distribution of the modern cephalopods is likely to be the presence/absence of the suitable food sources and thus the appearance of Arctic ammonites could be explained by presence of food accompanied by decreasing of diversity and competition. Cooling confined to Arctic and Boreal regions may similarly displace faunas equatorwards leading to an influx of Arctic/Boreal elements without necessarily a marked temperature drop in the Russian Platform.

6. Conclusions

The oxygen isotope data are supportive of a rapid rise of temperatures during the Oxfordian–Kimmeridgian. The magnitude of isotopic change (~4‰) is, however, less when compared to a number of belemnite-derived compilations of Jurassic isotope data. Hence previous large conjectured temperature shifts during this time interval may partly be an artefact of comparing data from dispersed locations. Although the late Kimmeridgian–Volgian have been identified as a period of time when temperatures reached a maximum, an episode of temperature stasis is indicated for this phase of the Jurassic of the Russian Platform.

The oxygen isotope data poorly record a number of significant oscillations in Boreal and Subtethyan ammonite dominant assemblages. This may suggest that belemnite derived isotopic data are not reflecting regional temperature trends but are instead dominated by localised variation of factors such as the $\delta^{18}O_{seawater}$. The reproducibility of trends across the Russian Platform perhaps negates such an interpretation. Alternatively temperature may not be a straightforward control upon cephalopod migration with other factors being more significant in controlling migration routes and destinations (e.g. Price et al., 2000; Wierzbowski, 2002).

Maximum $\delta^{13}C$ values are observed within the Volgian although the organic rich units of Gorodischi do not fall within this segment of the $\delta^{13}C$ curve. A positive carbon isotope excursion is not recorded within composite carbon-isotope stratigraphies of the Western Tethys. Reduced connections with the world ocean certainly reduce the possibility of recording global events. Hence the Jurassic may have been characterised by regional $\delta^{13}C$ excursions related to non-simultaneous organic matter deposition (e.g. Hantzpergue et al., 1998) resulting from localised ponding semi restricted ocean circulation and a lack of tidal mixing contrary to the Cretaceous period, where most of organic matter-rich deposits can be assigned to global oceanic anoxic events (e.g. Arthur et al., 1990).

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