Bio- and chemostratigraphic assessment of carbon isotope records across the Triassic–Jurassic boundary at Csővár quarry (Hungary) and Kendlbachgraben (Austria) and implications for global correlations

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Carbon isotope trends are useful for stratigraphic correlation, especially for time intervals when major perturbations of the global carbon cycle occurred. Such perturbations have been documented for the Triassic–Jurassic (T–J) boundary, and several successions from this time interval are charac-terized by (1) an initial negative excursion, followed by (2) a pronounced positive excursion and a subsequent (3) main negative carbon isotope excursion. These features, however, are not present in all T–J boundary sections, or the stratigraphic position of the positive or the main negative excursion has variable locations. In the present study, we analysed carbon isotopes in bulk carbonate from the pelagic Csővár quarry section in Hungary and from the intra-platform basin to shallow subtidal marine Kendlbachgraben section in Austria. Both T–J boundary successions are biostratigraphically well controlled enabling – with particular focus on the bio- and chemostratigraphy of other T–J boundary sections – correlation of the carbon isotope trends. This evaluation shows that the apex of the initial negative $\delta^{13}C$ excursion occurred slightly, but distinctly, below the mass extinction event and represents an excellent stratigraphic correlation tool.

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At the Triassic–Jurassic (T–J) transition, major environmental changes took place on Earth resulting in a major mass extinction event and distinct perturbations of oceanic and atmospheric geochemistry (e.g. Weems 1992; Hallam & Wignall 1997; Pálfy *et al.* 2001; Cohen & Coe 2002, 2007; Hesselbo *et al.* 2002; Pálfy 2003; Tanner *et al.* 2004; McElwain & Punyasena 2007; Hautmann *et al.* 2008; Kiessling *et al.* 2009; Kiessling 2010; Bonis 2010; Ruhl 2010; Ruhl *et al.* 2010a). Particularly large amplitude carbon isotope excursions have been reported for marine carbonate and organic matter across the T–J boundary (e.g. McRoberts *et al.* 1997; McElwain *et al.* 1999; Pálfy *et al.* 2001; Ward *et al.* 2001; Hesselbo *et al.* 2002; Guex *et al.* 2004; Galli *et al.* 2005; Lucas *et al.* 2007), and additional distinctive shifts have been also reported for the time span before and after the T–J boundary (Cleveland *et al.* 2008; Schaller *et al.* 2011; Ruhl & Kürschner 2011; Korte & Hesselbo 2011; Mette *et al.* unpublished manuscript).

Recent studies have shown that the geometry of the $\delta^{13}C$ trend across the T–J boundary is similar in many instances, consisting of: (1) an 'initial' negative excursion, followed by (2) a pronounced positive excursion, and (3) a 'main' negative excursion (Hesselbo *et al.* 2002; Ward *et al.* 2007; Kuerschner *et al.* 2007; Williford *et al.* 2007; McRoberts *et al.* 2007; Ruhl *et al.* 2009). A similar trend is also reported for terrestrial sediments (Hesselbo *et al.* 2002; Pieńkowski *et al.* 2011; Steinthorsdottir *et al.* 2011). These features were reported for bulk organic carbon as well as for well preserved low-Mg-calcite oysters which are resistant to diagenetic alteration (Korte *et al.* 2009). However, the 'main' negative excursion is not observed in all marine sections (e.g. see Pálfy *et al.* 2007; Galli *et al.*

2007; van de Schootbrugge *et al.* 2008; Götz *et al.* 2009), suggesting that local effects might have influenced the seawater $\delta^{13}C$ in some cases. The initial negative excursion, on the other hand, which is present in nearly all carbon isotope records of the discussed time interval, is large in magnitude (~ 5 ‰) for total organic carbon (TOC) in the Northern Alps sections (Kuhjoch and Hochalpengraben), but this magnitude is even larger (~8.5 ‰) when analysing specific biomarkers (Ruhl *et al.* 2011). This suggests that the carbon cycle was even more perturbed than previously thought.

The occurrence and the geometric similarities of the carbon isotope sequence in several localities of the Tethys and the region of western Pangaea argue for a secular $\delta^{13}C$ signal. Such large-scale parallel fluctuations are generally caused by perturbations in the Earth's carbon cycle (e.g. Kump & Arthur 1999, see also Beerling & Berner 2002), thus enabling the use of these carbon isotope excursions for stratigraphic correlation as demonstrated for many stage boundaries including the Permian–Triassic boundary (e.g. Kraus *et al.* 2009; Korte & Kozur 2010), the Cenomanian–Turonian boundary (e.g. Sageman *et al.* 2006), and for the base of the Eocene (e.g. Magioncalda *et al.*, 2004). $\delta^{13}C$ fluctuations across the T–J boundary have been used for chemostratigraphic correlation (e.g. McRoberts *et al.* 2007). In some cases, however, it is difficult to compare the organic and carbonate $\delta^{13}C$ excursions because shifts in bulk organic carbon isotopes can be caused by changes in the photosynthetic fractionation within plants due to modification of CO_{2aq}, influencing the growth rate of organisms (e.g. Kump & Arthur 1999). In addition, it has to be taken into account that the sources for carbonate and organic carbon isotopes might have been decoupled in the earliest Jurassic (Clémence *et al.* 2010). Because of all these factors it cannot be assumed that a simple one-to-one phase correlation can be expected for $\delta^{13}C_{org}$ and $\delta^{13}C_{carb}$, although it can be assumed that shifts go in the same directions.

Here we present new bulk carbonate $\delta^{13}C$ values for the T–J boundary sections at Csővár ('Csővár quarry', Cserhát Mts., Hungary) and Kendlbachgraben (Northern Alps, Austria) and make comparisons to carbon isotope results of bulk organic matter and pristine oyster-shells from the literature. The goal of the present study is to use the Csővár quarry conodont biostratigraphy and carbon isotopes to correlate to bio- and chemostratigraphical results from other sections. This correlation enables definition of a general carbon isotope trend of at least the initial negative excursion in high resolution and provides evidence for its outstanding correlation potential for a well-dated stratigraphic level somewhat below the T–J boundary.

Sampling localities

Csővár

Bulk rock carbonate samples originate from the Csővár quarry (also named Pokol-völgy quarry), northern Hungary (Fig. 1). This succession belongs to the Transdanubian Range Unit, representing – according to Haas *et al.* (1997) – the most internal part of the Transdanubian Range segment of the Tethys shelf. It comprises two limestone formations. The Tuvalian to Rhaetian Csővár Limestone Formation (Balogh 1981) consists of dark, bituminous limestones, partly graded (often with transported shallow water fossils and clasts from adjacent reefs), cherty limestones, and marls (all basinal to toe-of-slope sediments). The uppermost Rhaetian, Hettangian and Sinemurian Várhegy Cherty Limestone Formation consists of bedded, light-yellowish to light-brownish micritic limestones and cherty limestones (Kozur & Mock 1991; Kozur 1993). These two formations are regarded by other workers as a single formation (Csővár Limestone Formation *s.l.*) (e.g. Haas *et al.* 1997; Haas & Tardy-Filácz 2004). The investigated part of the succession comprises the Rhaetian and Hettangian stages and the *Misikella posthernsteini*, *Misikella ultima* and *Neohindeodella detrei* conodont zones (Fig. 2). The rich occurrence of Nassellaria-dominated radiolarians indicates deep water, but in some beds transported shallow-water reef fossils from adjacent reefs dominate (including corals, Kozur & Mostler 1973; Haas *et al.* 1997; Haas & Tardy-Filácz 2004). Especially striking is a thick bed at the top of the Csővár Limestone Formation *ss*. It contains numerous transported shallow-water fossils, among them also corals that were transported from adjacent reefs. Transported lithoclasts are also common. It seems that this bed indicates a sea-level drop that exposed the reefs and brought the new reef margin closer to the deposition area of the Csővár section. The maximum of this sea-level drop coincides with the extinction event of the Triassic faunal elements.

Immediately above these beds a sequence of marls, marly limestones, and silty shales begins, into which no shallow-water macrofossils and lithoclasts were transported. This is interpreted to indicate a sudden sea-level rise. Compared with the underlying beds, these marly, partly silty beds are very poor in fossils. Moreover, the carbonate production apparently decreased for a short interval (about 2 m in the section) and terrigenous input consists not only of clay, but also silt. This coupling of sea-level drop and immediately following sea-level rise confirms the scenario of an extinction event caused by cooling and contemporaneous sea-level drop, followed by warming (Schoene *et al.* 2010). This is also indicated by the extinction mode of conodonts. *Misikella ultima* disappeared first; this species was dominant before the extinction and occurs only in equatorial, very warm water environments. During the slightly younger main extinction event the warm water genus *Misikella* disappeared, whereas *Neohindeodella* continued, a genus which occurs both in warmer and cooler water.

The Csővár quarry beds (and the entire Tuvalian to Sinemurian basinal to toe-of-slope succession) were for a considerable period assigned to the lower Carnian. A latest Triassic age for the Csővár quarry locality was first reported by Kozur & Mostler (1973) and confirmed by Detre *et al.* (1986). Subsequently, Kozur & Mock (1991) recorded *M. ultima* and *N. detrei* from this quarry and defined the late Rhaetian *M. ultima* and *N. detrei* zones (Fig. 3), although for the latter zone it was not clear whether it extended into the Jurassic. A Hettangian age for the upper part of the *N. detrei* Zone was demonstrated by Kozur (1993) using radiolarians.

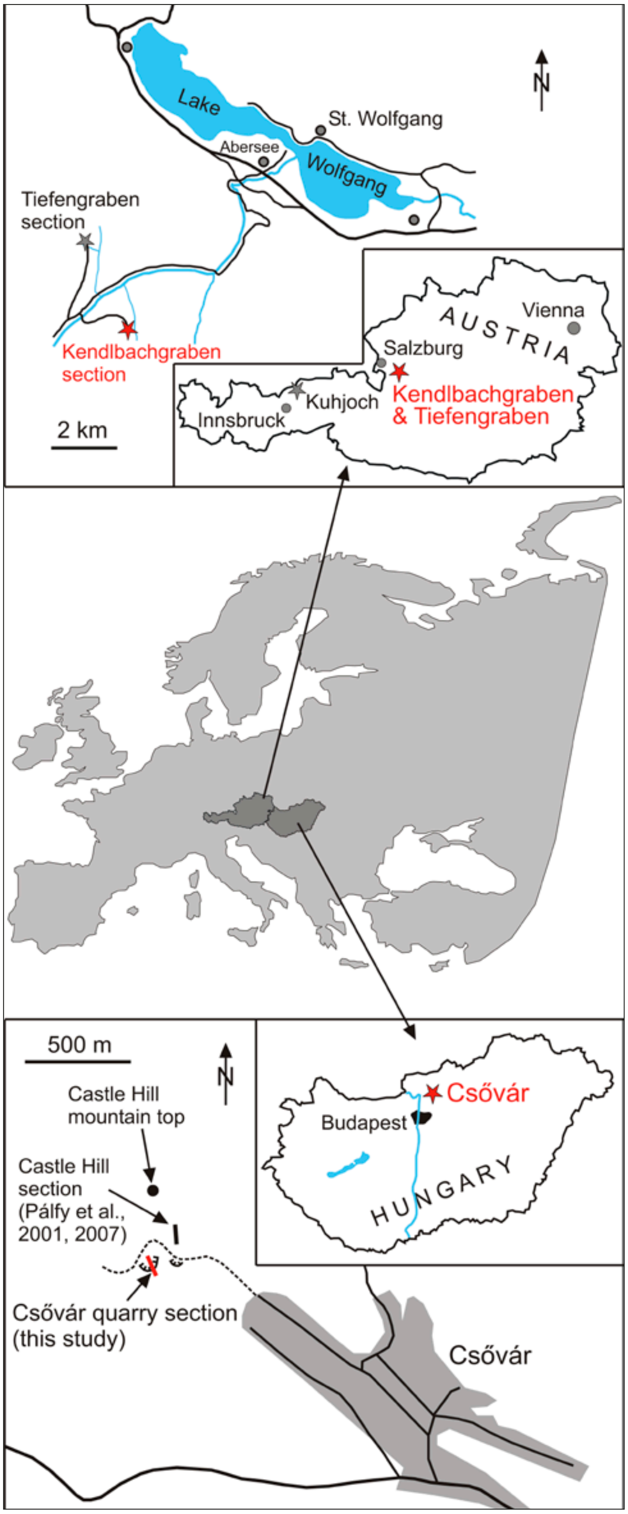


Fig. 1. Location maps, modified after Delecat (2005) and Bonis *et al.* (2009) for Kendlbachgraben (47° 41′ 15″ N / 13° 21′ 3″ E; Krystyn *et al.* 2005), as well as for the Csővár quarry (47° 49′ 09″ N / 19° 18′ 24″ E), including the location of the Castle Hill section at Csővár which was investigated by Pálfy *et al.* (2001, 2007).

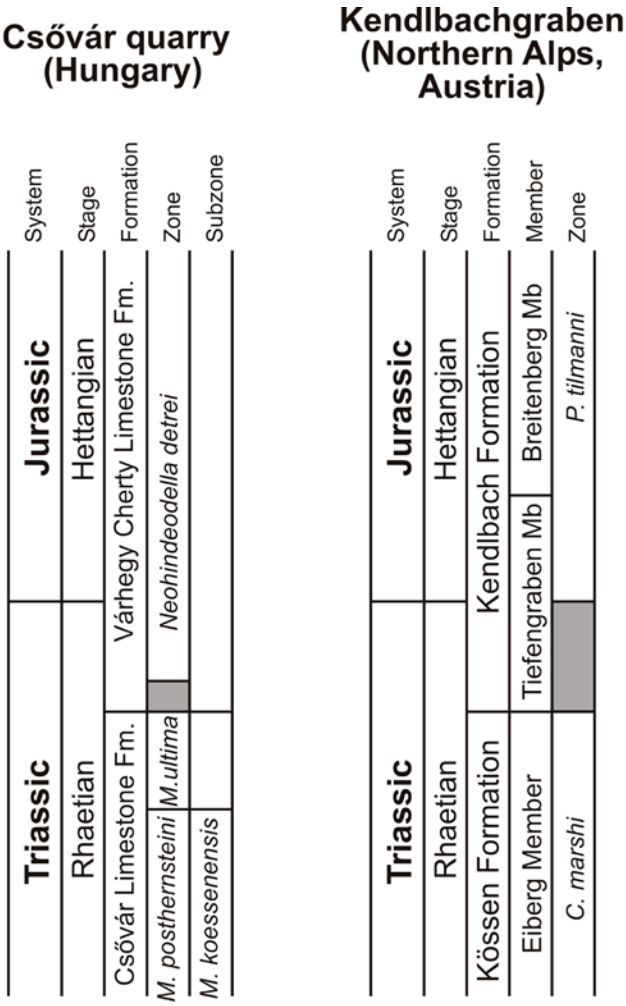


Fig. 2. Lithostratigraphic units and biostratigraphy of the investigated sections after Kozur & Mock (1991) and Kozur (1993) for Csővár and after Golebiowski (1990) and Kuerschner *et al.* (2007) for Kendlbachgraben.

Carbonates of the Csővár quarry are rich in conodonts; reworking of these (even in distal calciturbites) has not occurred, and the first appearance datum (FAD) and last occurrence datum (LOD) are easy to locate, well known, and possible to correlate with the isotope trends. Previously published isotope data are sparse (five ^{87}Sr/^{86}Sr ratios for conodonts (Korte *et al.* 2003) and only six δ^{13}C data for bulk carbonates (Korte *et al.* 2005)) and are restricted to the *M. posthernsteini* and *M. ultima* zones, and based on this sparse δ^{13}C dataset it was impossible to define an accurate carbon isotope trend.

Bulk carbonate δ^{13}C data across the T–J boundary exist from the Castle Hill (Vár-hegy) section at Csővár (Pálfy *et al.* 2001, 2007; Götz *et al.* 2009), a locality that is situated about 200 m north-east of the Csővár quarry (Fig. 1). A bed-by-bed correlation between the Castle Hill section and Csővár quarry is difficult because in the Castle Hill T–J boundary succession fewer conodonts can be found, and sedimentological complications (slump structures) occur. Calciturbites occur in both sections, but they only indicate transport of contemporaneous shallow-water deposits and slope material from the reef and reef slope into the basin; they do not contain any older conodonts. The slump structures in the Castle Hill T–J boundary succession may have resulted in somewhat older material being admixed into the succession. H.W. Kozur and R. Mock also investigated the conodont succession of the Castle Hill T–J boundary succession. As some reworked older conodonts were found, these data were not published and only the conodont data from the quarry section were published in Kozur & Mock (1991). As the aim of this work is exact dating of the initial negative excursion, we have restricted our investigations to the quarry section.

Kendlbachgraben

Bulk carbonate samples from the Kendlbachgraben section at the Lake Wolfgang (Austria) were investigated for isotopic composition (Fig. 4). The sediments were deposited in an intraplatform basin (Eiberg Basin) (Kuerschner *et al.* 2007). The succession comprises the Eiberg Member (higher part of the Kössen Formation) and the Tiefengraben and Breitenberg Members of the Kendlbach Formation. The *C. marshi* and the *P. tilmanni* ammonoid zones are present (Figs 2, 4). The marls of the lower Tiefengraben Member (Fig. 4) have earlier been named Grenzmergel (e.g. Hallam & Goodfellow 1990).

From the Kendlbachgraben locality, the Kössen and Kendlbachgraben Formations have been intensely studied biostratigraphically (e.g. Suess & Mojsisovics 1868; Morbey 1975; Krystyn 1980). Subsequently, Golebiowski & Braunstein (1988) have presented detailed biostratigraphic fossil ranges in comparison to those of the Tiefengraben section; the latter locality is situated approximately 3 km north of the Kendlbachgraben section (Fig. 1). For Kendlbachgraben, δ^{13}C was analysed in bulk carbonate by Hallam & Goodfellow (1990), but Morante & Hallam (1996) suggested that the ^{13}C-depleted data of these authors, at least for the mudstones, are most probably diagenetically altered because a distinct positive δ^{13}C excursion occurred coevally for bulk organics. A similar positive trend in the mudstones was reported for the same section

Table 1. Analytical data. Samples are numbered by their relative height in metres in the sections in Figs 3 and 4.

Csővár			Kendlbachgraben		
m	δ^{13}C	δ^{18}O	m	δ^{13}C	δ^{18}O
-12.3	2.20	-1.51	-355	3.00	-1.27
-11.6	2.95	-2.75	-337	2.73	-1.73
-11.1	2.26	-0.79	-311	3.05	-1.36
-11.0	2.13	-0.64	-283	2.94	-1.25
-10.8	2.00	-0.44	-275	2.88	-1.37
-10.5	2.24	-0.97	-244	2.87	-1.90
-10.4	2.12	-0.64	-211	2.88	-2.65
-10.0	2.12	-1.20	-177	2.79	-2.02
-9.7	2.05	-0.71	-145	2.67	-1.44
-9.1	2.46	-0.34	-118	2.41	-1.71
-8.8	2.26	-1.38	-93	2.59	-2.68
-8.2	2.44	-0.86	-76	2.42	-2.25
-6.0	2.81	0.29	-73	2.40	-1.99
-2.0	1.52	-6.67	-56	2.27	-3.81
-1.5	1.88	-9.68	-54	2.61	-2.31
-1.2	1.36	-1.74	-49	2.56	-2.09
-0.9	1.75	-3.58	-22	2.51	-2.14
-0.6	1.44	-5.76	-6	2.29	-1.21
-0.3	2.26	-4.74	-5	2.32	-0.79
0.0	2.17	-5.13	-1	1.80	-1.99
0.1	2.43	0.20	305	2.81	-1.17
0.2	1.68	-4.09	305	2.84	-1.29
0.4	2.26	-3.04	318	2.81	-0.76
0.7	2.03	-13.06	318	2.87	-1.10
0.9	2.31	-1.82			
1.0	2.06	-2.48			
1.1	2.45	-1.88			
1.4	2.65	-2.05			
1.7	2.72	-1.83			
2.0	2.79	-1.64			
5.0	2.39	-1.66			
5.6	3.50	-1.41			
6.3	3.01	-5.19			

by Ruhl *et al.* (2009), for the Tiefengraben section by Kuerschner *et al.* (2007), and for other sections in the Northern Alps (Ruhl *et al.* 2009).

Material and methods

Bulk rock carbonate from both sections was analysed for stable isotopes. Samples of 2 to 5 mg of fine-grained bulk carbonates – drilled from fresh surfaces – were analysed for $\delta^{13}C$ and $\delta^{18}O$ at the Department of Earth Sciences at the University of Oxford using the VG Isogas Prism II mass spectrometer with an online VG Isocarb common acid bath preparation system. The samples reacted in the instrument with purified phosphoric acid (H_3PO_4) at 90°C. Calibration to the V-PDB standard via NBS-19 is made daily using the Oxford in-house (NOCZ) Carrara Marble standard. Reproducibility of replicated standards was better than 0.1‰ for both carbon and oxygen isotopes. $\delta^{13}C$ and $\delta^{18}O$ values (Table 1) have been reported in delta notation relative to the Vienna Pee Dee Belemnite (V-PDB) international scale.

Biostratigraphy was investigated in the field and by the determination of conodonts (acetic acid treatment). The positions of the analysed samples are shown in Figs 3 and 4.

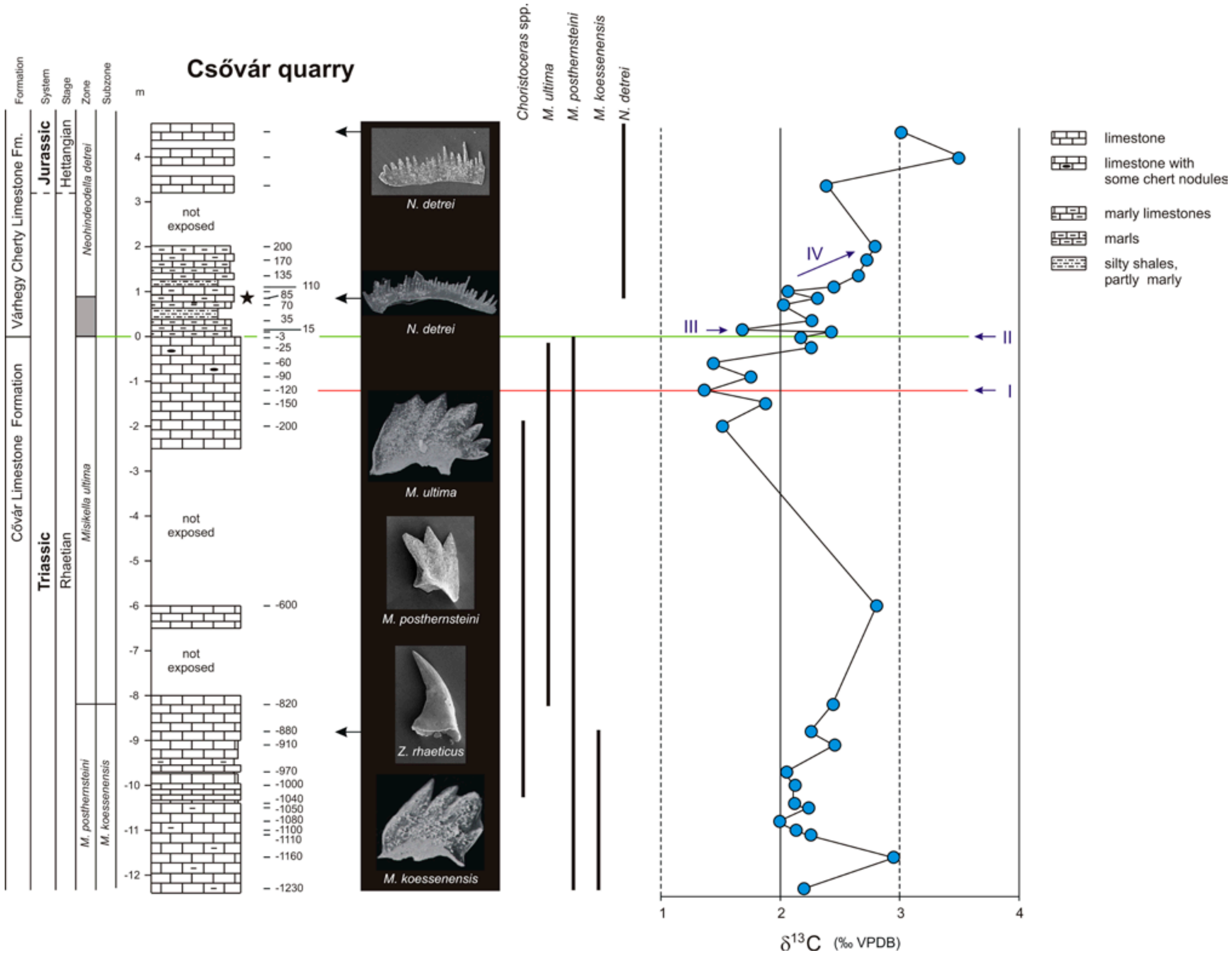


Fig. 3. Csővár quarry section, showing biostratigraphy, sample locations and carbon isotope values for bulk carbonates. The stratigraphically most important late Rhaetian and basal Hettangian conodonts are shown. Star: level of the transitional ammonoid form (see text and Fig. 5); Red line: apex of the initial negative carbon isotope excursion; Green line: LOD of Triassic *Misikella* conodonts; Grey shadow: 'pre-*planorbis* Beds' without definitive Triassic and Jurassic faunas; (I) minimum of the initial negative $\delta^{13}C$ excursion; (II) LOD of *Misikella* and other Triassic conodonts and beginning of the 'grey zone' ('pre-*planorbis* Beds'); (III) second negative $\delta^{13}C$ excursion in the basal part of the 'grey zone' ('pre-*planorbis* Beds'); (IV) a positive carbon isotope excursion.

Results

Biostratigraphy

The new conodont biostratigraphic investigations at Csővár quarry confirm the results of Kozur & Mock (1991) and Kozur (2003a, b). Most of the quarry section belongs to the late Rhaetian *M. ultima* conodont Zone (Fig. 3). Triassic conodonts remain common and diverse up to the top of this biozone and only *Norigondolella steinbergensis* (Mosher) disappears earlier. Ammonoids in the *M. ultima* Zone are rare and mostly represented by different *Choristoceras* species. *Choristoceras* is absent in the uppermost metre of the *M. ultima* Zone, but the remaining fauna remains abundant and diverse below the 'pre-*planorbis* Beds' for both pelagic faunal elements (especially badly preserved radiolarians of the late Rhaetian *Globolaxtorium tozeri* Zone) and transported shallow-water elements from adjacent reefs.

A major facies change occurs in the uppermost Csővár quarry section. The very fossil-rich, dark, bituminous, partly cherty limestones, which are often distinctly laminated especially in the upper part, are overlain with a sharp boundary by poorly fossiliferous, marly-silty beds with some marly limestone intercalations ('Pre-*planorbis* Beds'; grey zone in Fig. 3). An unconformity is not present at this level; the water remains deep, but a sea-level drop is indicated for the uppermost bed of the *M. ultima* Zone (see above). A

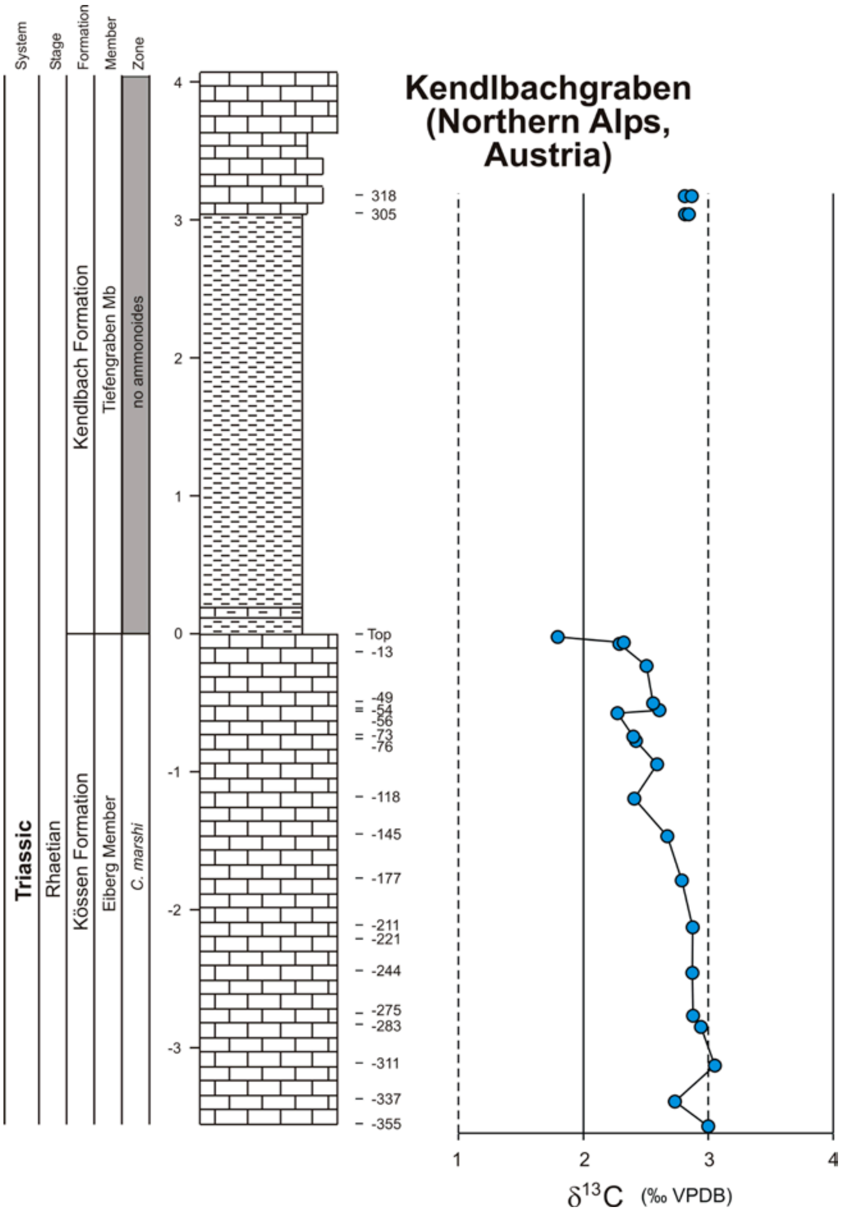


Fig. 4. Stratigraphic section for Kendlbachgraben, showing sample locations and carbon isotope values for bulk carbonates (see also the caption to Fig. 3).

distinct mass extinction of Triassic faunal elements (e.g., the rich *Misikella* conodont fauna), somewhat postdating the disappearance of the *Choristoceras* ammonoid fauna, occurs at the facies boundary. The sediments above the sharp lithological facies change are named the 'pre-*planorbis* Beds' of the basal Várhegy Cherty Limestone Formation; a very poor conodont fauna of the *N. detrei* Zone occurs in these deposits consisting of only very few representatives of the index species. This taxon, *N. detrei* (Kozur & Mock), is somewhat more common in the overlying micritic limestones (Fig. 3) where it continues to be found up to beds with poorly preserved earliest Jurassic radiolarians and Psiloceratids indet. (Kozur 1993; Götz *et al.* 2009). This occurrence of *N. detrei* represents the youngest (early Hettangian) conodont fauna of the world (Kozur 2003a, b).

A single ammonoid specimen has been recovered from a marly limestone intercalation about 0.9 m above the base of the 'pre-*planorbis* Beds' of the Csövár quarry section (Fig. 3: star, Fig. 5). This ammonoid was determined by Prof. Jean Guex (Lausanne) as a transitional form between the Triassic *Rhacophyllites* and the Hettangian *Psiloceras*, confirming the transitional Rhaetian–Hettangian character of these beds. In the 'pre-*planorbis* Beds' and time-equivalents between the last *Misikella* and the FAD of *Psiloceras spelae* (Guex), ammonoids are virtually absent world-wide (e.g. in New York Canyon (USA) a short ammonoid-free interval is present between the uppermost Triassic *Choristoceras crickmayi* zone and lowermost Jurassic *Psiloceras tilmanni* zone ammonoid faunas). Hence, the discovery of this Rhaetian–Hettangian transition ammonoid emphasizes the importance of this locality for biostratigraphy.

Carbon isotopes

Bulk carbonate carbon isotope values from the Csővár quarry (Fig. 3) vary between 2 and 3 ‰ in the *Misikella koessenensis* subzone of the *M. posthernsteini* zone, and are about 2 ‰ at its top. The values increase to 2.8 ‰ from the lower to the middle *M. ultima* Zone followed by a decrease to values of about 1.5 ‰ about 1 m below the top of the *M. ultima* Zone. By using biostratigraphy as a correlative baseline, we conclude that these low δ^{13}C values in the upper *ultima* zone (Fig. 6) represent the initial negative carbon isotope excursion at the T–J boundary (cf. Hesselbo *et al.* 2002). In the Csővár quarry section it is a multiple signal (Fig. 3) consisting of three small negative shifts situated between 2 m and 0.6 m below the top of the *M. ultima* zone (around I in Fig. 3), separated by slight, short recoveries. After a distinct recovery to about 2.4 ‰ (II in Fig. 3), a fourth weaker negative signal occurs 0.15 m above the top of the *M. ultima* zone (III in Fig. 3) within the basal 'pre-*planorbis* Beds'. This negative shift 0.15 m above the top of the *M. ultima* zone could be interpreted as the last signal of the initial negative shift at the T–J boundary. Instead, we suggest that this represents an additional traceable smaller negative excursion that is also observable at the same stratigraphic level at Lavernock Point, St. Audrie's Bay, Tiefengraben and New York Canyon (Fig. 6). This additional negative carbon isotope excursion is separated from the initial negative shift (cf. Hesselbo *et al.* 2002; Ruhl *et al.* 2009) by a distinct δ^{13}C recovery (just below the green line in Figs 3 and 6). Above this additional negative shift the curve increases to values of 3.5 ‰ in the *N. detrei* zone (IV in Fig. 3), representing the positive δ^{13}C excursion at the T–J boundary *sensu* Hesselbo *et al.* (2002).

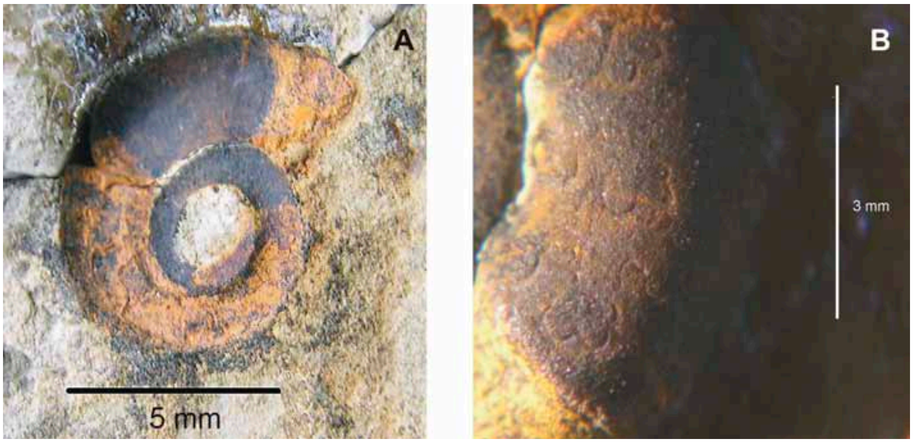
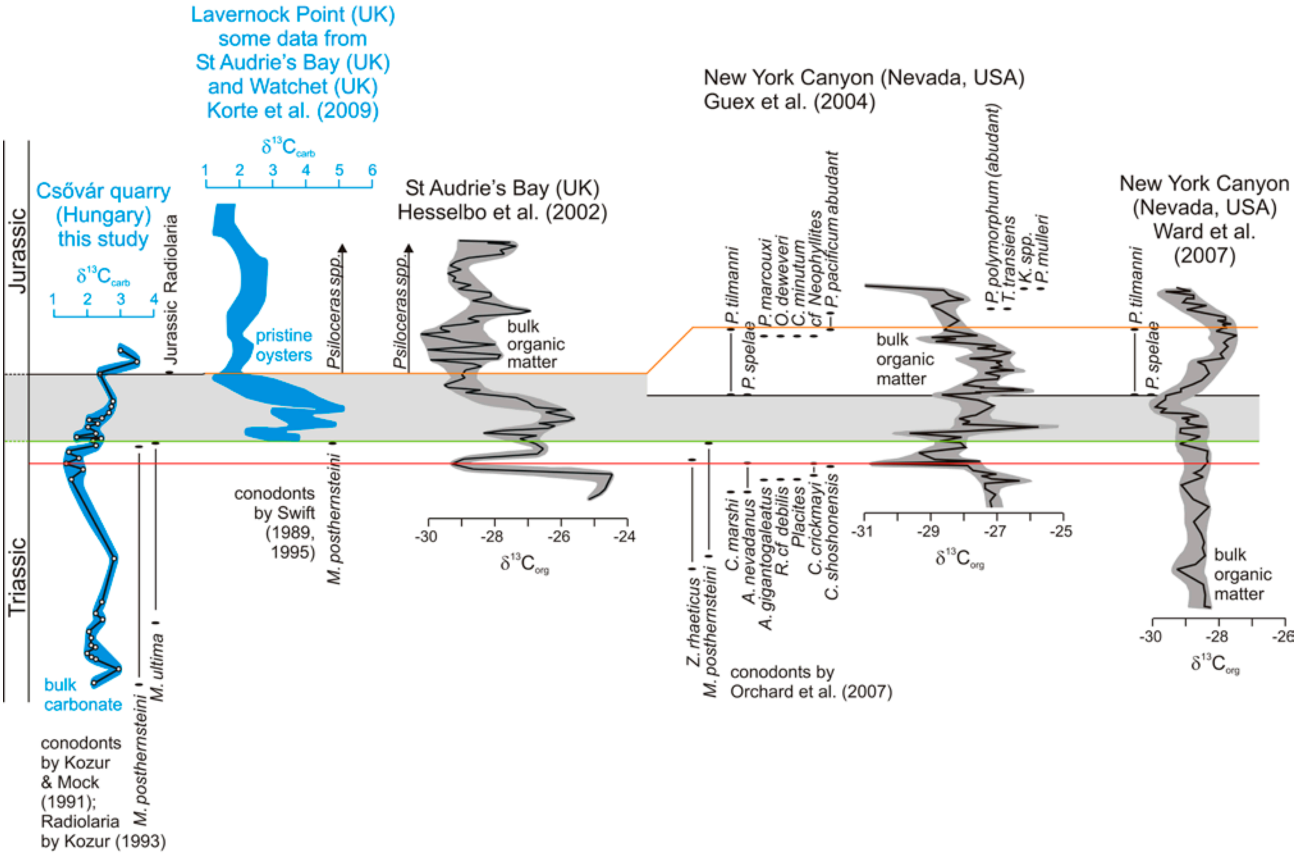


Fig. 5. Transitional form between the Triassic earliest Hettangian *Rhacophyllites* and the early Hettangian *Psiloceras* (det. Prof. Jean Guex, Lausanne), 0.9 m above the base of the 'pre-*planorbis* Beds' (= 0.9 m above the LOD of *M. posthernsteini* Kozur & Mock). This form has not been described previously. The level from which the ammonoid has been derived (above the last Triassic conodonts = above the LOD of *Misikella*, and below *Psiloceras tilmanni* (Lange) and *P. spelae* (Guex), in general has no ammonoids. a) entire specimen, b) detail with suture line. This specimen documents that real Jurassic ammonoids were not yet present after Triassic forms had disappeared.

The carbon isotope curve from the Kendlbachgraben section decreases gradually from about 3 ‰ to 1.8 ‰ in the *Choristoceras marshi* zone and the lowest value is situated at the top of the Kössen beds (Fig. 4). For the overlying Kendlbach Formation no data were obtained for the mudstones and δ^{13}C values of nearly 3 ‰ are obtained from the base of the overlying limestones (see also Morante & Hallam 1996).

Discussion

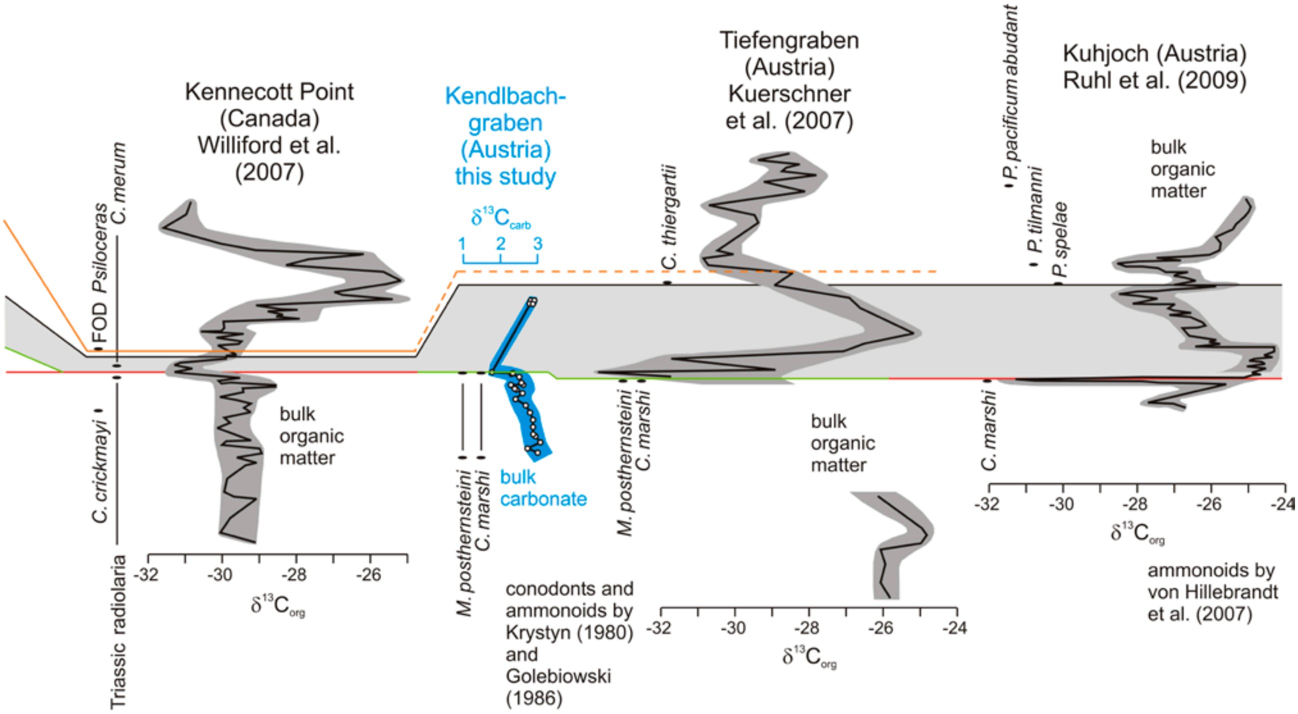
To be able to utilize the T–J boundary carbon isotope trends for stratigraphic correlation it is necessary to constrain the δ^{13}C fluctuations biostratigraphically. For this purpose, the Csővár quarry is important because of the presence of the well known FAD and LOD of conodonts. The importance of the carbon isotope trend from the Csővár quarry becomes clear when it is compared to bulk organic and pristine oyster-shell data from the literature (Fig. 6). The initial δ^{13}C minimum at Csővár quarry is about 1 m (between 0.6 to 2 m) below the top of the *ultima* zone. This result is similar to that for the Csővár Castle Hill section (see Pálfy *et al.* 2001, 2007; Haas & Tardy-Filácz 2004; Götz *et al.* 2009; Haas *et al.* 2010), but we note that the data by Pálfy *et al.* (2001, 2007) vary between – 4 and + 2 ‰ in their T–J boundary interval, whereas the T–J boundary interval data of the present study (~ 'grey zone') are between +1.5 and +2.5 ‰ (Fig. 3). At St Audrie's Bay, the initial negative carbon isotope excursion in the Lilstock Formation is about 1 m below the LOD of *M. posthernsteini* discovered in another nearby section (Fig. 6; see also Swift 1989, 1995; Hesselbo *et al.* 2002). The carbon isotope values from the New York Canyon (Ferguson Hill) section (Nevada, USA) are also important. Ranges for ammonites (Guex *et al.* 2004, Lucas *et al.* 2007) and conodonts (Orchard *et al.* 2007) from this locality are well established. Several δ^{13}C fluctuations and a distinct initial negative excursion are observed in the carbon isotope values of this locality (Guex *et*



▲▶Fig. 6: Carbon isotope trends of bulk carbonates, pristine oysters and bulk organics for different T–J boundary sections, for comparison. Grey shadow: Interval without definitive Triassic and Jurassic faunas. Red line: apex of the initial negative carbon isotope excursion. Green line: LOD of Triassic *Misikella* conodonts. Orange line: FAD of moderately advanced *Psiloceras*, such as *Psiloceras planorbis* (Sowerby), *Psiloceras pacificum* (Guex). The correlation documents the chemostratigraphic correlation potential of the initial negative carbon isotope excursion. (Figure developed after McRoberts *et al.* 2007).

al. 2004; see Fig. 6). A different carbon isotope feature is reported by Ward *et al.* (2007) for the same section: a distinct initial negative shift is not recorded and a positive excursion occurs stratigraphically higher (Fig. 6; see also Guex *et al.* 2004). However, the first negative carbon isotope excursion at the New York Canyon recorded by Guex *et al.* (2004) is about 2 to 2.5 m below the LOD of *M. posthernsteini* (see Orchard *et al.* 2007 for conodont distribution) and this characteristic, as well as its shape and magnitude, is very similar to the initial negative carbon isotope excursion at St Audrie's Bay (Hesselbo *et al.* 2002).

Accepting that the negative New York Canyon excursion about 2 m below LOD of *M. posthernsteini* recorded by Guex *et al.* (2004, see also Lucas *et al.* 2007) is a reliable isotope signature, then we have a correlateable, uniform, conodont-controlled stratigraphic level for the initial T–J boundary δ^{13}C minimum at three different regions in Hungary, England and Nevada. These δ^{13}C minima are all distinctly (0.6 to 2.5 m) below the disappearance of the genus *Misikella*, marked by the red line in Fig. 6. At this point it is necessary to explain the distribution of the genus *Misikella*. *Misikella* occurred only at low latitudes; *M. ultima* was restricted to equatorial regions, whereas *M. posthernsteini* was distributed somewhat farther to the north (up to the Rhaetian palaeolatitudes of England and Nevada) compared to *M. ultima*. In equatorial regions (e.g. Hungary), *M. posthernsteini* and *M. ultima* became extinct almost contemporaneously. Therefore, the LOD of *M. ultima* and *M. posthernsteini* in Hungary and in the Alps corresponds to the LOD of *M. posthernsteini* in England and Nevada and defines the disappearance of the latest Triassic-type conodont faunas. At the Kennecott Point section (Queen Charlotte Islands, Canada), *Misikella* does not occur because this locality was situated even farther to the north than England and Nevada in T–J times, but here the disappearance of *Norigondolella* corresponds approximately to the LOD of *Misikella* and this is about 2–3 m below the first appearance of Jurassic radiolarians (M. Orchard, personal communication 2010).



A different feature is visible for the Alpine Kendlbachgraben and Tiefengraben sections (Fig. 6). Here the disappearance of *M. posthernsteini* and of *Choristoceras marshi* (Hauer) are contemporaneous (Fig. 6; see also Kuerschner *et al.* 2007) whereas in other sections *C. marshi* disappears distinctly before the LOD of *M. posthernsteini* and *M. ultima*. The contemporaneous disappearance of *C. marshi* and *Misikella* at the boundary between the Kössen and Kendlbach formations at a negative carbon isotope shift indicates either a short stratigraphic gap immediately before the 'pre-*planorbis* Beds', or a major condensation of the sequence. In all low-latitude sections towards the north as far as New York Canyon and St Audrie's Bay, the initial negative excursion lies distinctly below the LOD of *Misikella*. This observation is confirmed by von Hillebrandt *et al.* (2007) who pointed out that the equivalent of the uppermost Kössen Formation Bed T, of around 20 cm thickness in the GSSP Kuhjoch section, Austria, is absent at Kendlbachgraben and Tiefengraben due to non-deposition.

No conodonts exist for the uppermost Kössen Formation at the Kuhjoch section (Fig. 6). Therefore, it is not biostratigraphically clear whether a gap is present in the Kuhjoch section above bed T of the topmost Kössen Formation, as observed at Kendlbachgraben and Tiefengraben. However, the negative δ^{13}C peak in the Kuhjoch section is situated at the boundary between the Kössen Formation and the Kendlbach Formation (Tiefengraben Member), immediately above a limestone unit, similar to the relationships observed at Kendlbachgraben and Tiefengraben. The stratigraphic position of this δ^{13}C negative peak at Kendlbachgraben and Tiefengraben corresponds most likely to the negative shift in the basal part of the 'pre-*planorbis* Beds' in the Csővár quarry section and not to that of the upper *M. ultima* Zone. On the other hand, the LOD of *C. marshi* at the level of the negative excursion in the Northern Alps sections indicates that this negative shift corresponds to the initial negative shift. In both cases, a short gap must be present at Kendlbachgraben and Tiefengraben, arguing for a condensation or a non-sedimentation level in the basal part of the 'pre-*planorbis* Beds'. An additional correlation marker for the initial negative shift is a peak in prasinophytes that can be observed in the Alpine sections and Csővár at the level of the upper part of the initial negative excursion in the basal Kendlbach Formation of the Eiberg Basin in the Northern Alps and in the uppermost Csővár Limestone Formation in Hungary (Bonis *et al.* 2009, 2010; Götz *et al.* 2009; Bonis 2010; Ruhl *et al.* 2010b). In addition to the stratigraphic importance of the prasinophyte maximum during the (upper part of the) initial negative excursion, these fossils have also a palaeoecological importance. These 'disaster species' can be often found close to levels of mass extinctions. Otherwise, they either prefer brackish to freshwater conditions or low water temperatures (e.g. Prauss 2007; Bonis 2010). Bonis *et al.* (2010) favoured brackish conditions because no glaciations are known close to the T–J boundary. However, shorter cooling events will not lead to glaciations, and cooler water temperatures at low latitudes can also be present in times without high latitude glaciations. Moreover, in partly restricted basins such as the Eiberg Basin, brackish conditions can easily be established. However, at Csővár, open sea marine conditions prevailed at the level of the initial negative excursion and the prasinophyte acme. Here, somewhat lower water temperatures are favoured because brackish conditions can be excluded by the stenohaline marine fauna. Furthermore, the mode of the mass extinction instead indicates rapid cooling as the cause (see above). Such short periods of cooler temperatures are characteristic of large scale volcanism, such as the Central Atlantic Magmatic Province (CAMP) plateau basalts at the T–J boundary (Courtillot & Renne 2003). Also in the uppermost Permian, mass extinction events occur that are related to both global warming and global cooling events (Korte & Kozur 2011; Kozur & Weems 2011). Kozur & Weems (2005) have used conchostracans to show that the CAMP volcanism was not restricted to the Jurassic, but that the first lava flow in the Newark Supergroup belongs to the upper Rhaetian because it is immediately under- and overlain by late Rhaetian conchostracan faunas. Cirilli *et al.* (2009) have also reached similar conclusions by evaluation of sporomorphs. Thus, the main extinction event of marine uppermost Triassic low latitude faunas is apparently related to the onset of plateau basalt volcanism within the CAMP, just as the main extinction event within the uppermost Permian of marine low latitude faunas is contemporaneous with the onset of plateau basalt effusion above the tuffs within the Siberian Trap (Renne & Basu 1991; Korte *et al.* 2010; Kozur & Weems 2010, 2011). The latest Permian low latitude marine main extinction event was suggested to be caused by global cooling, whereas the earlier continental and high latitude marine extinction horizon within the uppermost Permian was caused, in contrast, by global warming (Kozur & Weems 2011). Such different phases of short-lasting global cooling and longer-lasting global warming may also be present within the uppermost Triassic and affected the biota (Schoene *et al.*, 2010).

As a consequence of the above discussion, an essential correlation horizon around the low latitude T–J boundary is the LOD of *Misikella* (*M. posthernsteini, M. ultima*) and this is clearly seen in the uncondensed sections at Csővár and New York Canyon to occur above the disappearance of Triassic ammonoid species of

Choristoceras (*C. marshi, C. crickmayi*) (Fig. 6). The initial δ^{13}C minimum is coeval with the disappearance of *C. marshi* and *C. crickmayi*. Note that *C. crickmayi* may occur stratigraphically somewhat higher up than *C. marshi*, but this phenomenon might be a LOD of the latter species in Nevada because *C. crickmayi* is a typical North America form, but *C. marshi* is not.

The initial T–J boundary δ^{13}C minimum is synchronous; its position occurs near the LOD of *C. crickmayi* (and at or somewhat above the LOD of *C. marshi*) and somewhat below the disappearance of genus *Misikella*.

A second time line is the base of the 'grey zone' (green line in Fig. 6) at the LOD of *Misikella*, and this is somewhat above the initial δ^{13}C minimum. In the 'grey zone', faunas are not definitively Triassic or Jurassic. Significantly, the transitional form between the Triassic ammonoid *Rhacophyllites* and the Hettangian ammonoid *Psiloceras* was found in this horizon at the Csővár quarry section. Both the red and the green time lines contrast for the Alpine localities because, for Kendlbachgraben, Tiefengraben and Kuhjoch, the base of the 'grey zone' starts already at the initial δ^{13}C minimum, and even somewhat lower and somewhat below the δ^{13}C minimum at Tiefengraben. We also note that in Kössen intraplatform basins the disappearance of *Misikella* occurs at the top of the Koessen limestone (see Golebiowski & Braunstein 1988) and there the 'pre-*planorbis* Beds' ('grey zone') begin at the initial δ^{13}C minimum. At Csővár, in England, and in New York Canyon, the 'pre-*planorbis* Beds' begin somewhat later, at the LOD of *Misikella* above the initial δ^{13}C minimum. Alternatively, and more probably, they begin at the same level but there is a short gap below the 'pre-*planorbis* Beds' in the Kendlbachgraben and Tiefengraben sections (see above), and perhaps also in the Kuhjoch section (no conodonts are known from this section).

The top of the 'grey zone' is diachronous and therefore not a timeline. The 'grey zone' ends with the first ammonite indicative of a Jurassic age, but we note that the first *Psiloceras* from the New York Canyon (Nevada) and Kuhjoch sections (Austria) are older than *P. planorbis*. The FAD of *Psiloceras* spp. at Ferguson Hill (see Guex *et al.* 2003, 2004; Lucas *et al.* 2007; Ward *et al.* 2007) is defined by the appearance of *P. tilmanni* (Lange) and *P. spelae* whereas at the Kuhjoch section it is defined by the FOD of *P. spelae tirolicum* Hillebandt & Krystyn (von Hillebrandt *et al.* 2007; von Hillebrandt & Krystyn 2009). The FAD of this latter subspecies may postdate the FAD of *P. spelae* in Nevada by up to 100 000 years (Ruhl 2010). In ammonoid-free marine sections and in continental sections, the FAD of the sporomorph *Cerebropollenites thiergartii* (Schulz) may be used as a proxy for correlation with the base of the *P. tilmanni* Zone, because this species begins only somewhat earlier in the Kuhjoch section (von Hillebrandt *et al.* 2007; Bonis *et al.* 2009, 2010; Bonis 2010), and Kuerschner *et al.* (2007) used the FAD of *C. thiergartii* for definition of the base of the Jurassic. Jurassic radiolarians seem to appear somewhat before the oldest *Psiloceras* (Williford *et al.* 2007).

The first *Psiloceras* in England belongs to the younger *P. planorbis*-group, rather than to *P. tilmanni* or *P. spelae*. The appearance of *Psiloceras* spp. (*Psiloceras planorbis* group) in England is in accordance with *P. marcouxi*, *P. pacificum*, *P. polymorphum* in Nevada and therefore a further timeline can be constructed (orange line in Fig. 6). The orange timeline is an important baseline for further considerations.

For the 'grey zone' a main positive carbon isotope excursion was reported by Hesselbo et al. (2002) followed by the main negative excursion just below the orange time line, and this general trend has been confirmed by Kuerschner et al. (2007), Korte et al. (2009) and Ruhl et al. (2009) as well as in the present study of pelagic carbonates in the Csővár quarry section (Figs 3 and 6). Albeit not as pronounced as in other sections, this general positive excursion is also recognizable in the New York Canyon dataset (Guex et al. 2004, 2006) (Fig. 6). For St Audries's Bay (Hesselbo et al. 2002), Lavernock Point (Korte et al. 2009), New York Canyon (Guex et al. 2004) and Tiefengraben, the first negative peak of the main negative excursion is slightly below the orange line; for the other sections not enough data exist to delineate a trend in this stratigraphic level. Contrasting features (Fig. 6) are reported for both the New York Canyon Ferguson Hill section (Ward et al. 2007) and for Kennecott Point (Williford et al. 2007). At these localities, a negative excursion occurs in the middle of the 'grey zone' and a distinct positive excursion is reported at or above the orange time-line; this distinct positive excursion at Ferguson Hill and Kennecott Point is at that level in which the first negative peak of the main negative excursion occurs in other sections (Fig. 6). A late Hettangian age for the positive carbon (Fig. 6) (and sulphur) isotope excursion(s) by Williford et al. (2007, 2009) for Kennecott Point is also suggested by further biostratigraphic evaluations by Guex et al. (2011). The main negative excursion in the Kuhjoch section begins considerably below the FAD of *P. spelae* and ends within the *P. tilmanni* Zone and cannot be correlated with any other section. In all other sections most of the main negative excursion lies much higher, mainly above the *P. tilmanni* Zone. Hence, the main negative excursion in the Kuhjoch section does not occur at the same time as the main negative excursion in St Audrie's Bay and other sections.

Conclusions

Integrated biostratigraphy (conodonts and ammonoids) and carbon isotope stratigraphy for the T–J boundary sections at Csővár quarry (Hungary) and comparison with fossil distribution and $\delta^{13}C$ trends of coeval localities (St Audrie's Bay, Lavernock Point, New York Canyon, Kennecott Point, Tiefengraben, Kuhjoch), allows a succession of stratigraphic markers to be defined. The apex of the initial negative carbon isotope excursion occurs approximately at the LOD of *C. crickmayi* and *C. marshi* (possibly somewhat above the latter), and slightly below (0.6 to 2 m) the disappearance of the genus *Misikella*. The positive excursion culminated in the 'grey zone', in which neither definitive Triassic nor definitive Jurassic faunas occur. The first negative peak of the main negative isotope excursion is most probably slightly below the timeline defined by the first appearance of *Psiloceras* spp. *(Psiloceras planorbis* group) in England and *P. marcouxi*, *P. pacificum*, *P. polymorphum* in Nevada. Therefore, the combined bio- and event chronology in ascending order is as follows:

1. *C. crickmayi* and *C. marshi* (the latter perhaps somewhat earlier) disappear contemporaneously with the occurrence of the initial negative $\delta^{13}C$ excursion minimum.
2. LOD of *Misikella* and other Triassic conodonts and beginning of the 'grey zone' ('pre-*planorbis* Beds').
3. A second negative $\delta^{13}C$ excursion (maybe a last minimum of the initial negative excursion) in the basal part of the 'grey zone' ('pre-*planorbis* Beds') insignificantly above the LOD of *Misikella* and other Triassic conodonts (at Csővár quarry section 0.15 m above the LOD of *Misikella*).
4. A positive $\delta^{13}C$ excursion (weak at Csővár and Kuhjoch sections, not well defined at New York Canyon (one high value by Guex *et al.* 2004).
5. FAD of primitive *Psiloceras* (*P. spelae*, *P. tilmanni*).
6. FAD of moderately advanced *Psiloceras* (such as *P. marcouxi*, *P. pacificum*, *P. planorbis*).
7. The main negative $\delta^{13}C$ excursion, beginning somewhat below event 6, but most of it lies above event 6. This excursion is not identical with main negative excursion at the Kuhjoch section.

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