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Chemostratigraphy Across the Triassic–Jurassic Boundary

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ABSTRACT

The Triassic-Jurassic transition (~201.5 Ma) is marked by one of the largest mass extinctions in Earth’s history. This was accompanied by significant perturbations in ocean and atmosphere geochemistry, including the global carbon cycle, as expressed by major fluctuations in carbon isotope ratios. Central Atlantic Magmatic Province (CAMP) volcanism triggered environmental changes and played a key role in this biotic crisis. Biostratigraphic and chronostratigraphic studies link the end-Triassic mass extinction with the early phases of CAMP volcanism, and notable mercury enrichments in geographically distributed marine and continental strata are shown to be coeval with the onset of the extrusive emplacement of CAMP. Sulfuric acid induced atmospheric aerosol clouds from subaerial CAMP volcanism can explain a brief, relatively cool seawater temperature pulse in the mid-paleolatitude Pan-European seaway across the T–J transition. The occurrence of CAMP-induced carbon degassing may explain the overall long-term shift toward much warmer conditions. The effect of CAMP volcanism on seawater ⁸⁷Sr/⁸⁶Sr values might have been indirect by driving enhanced continental weathering intensity. Changes in ocean-atmosphere geochemistry and associated (causative) effects on paleoclimatic, paleoenvironmental, and paleoceanographic conditions on local, regional, and global scales are however not yet fully constrained.

10.1. INTRODUCTION

The base of the Jurassic system, and therewith the Triassic-Jurassic (T–J) boundary (TJB) (201.36 Ma, Wotzlaw et al., 2014), is defined in the Global Boundary Stratotype Section and Point (GSSP) at Kuhjoch (Karwendel Mountains, Northern Calcareous Alps, Tyrol, Austria) [Hillebrandt et al., 2013]. The TJB transition at Kuhjoch and at correlated successions in the same area [see Hillebrandt et al., 2013, for overview] as well as many other coeval successions around the globe [see Hesselbo et al., 2007, for overview] are sedimentologically, paleontologically, and chemostratigraphically well studied (Fig. 10.1). It has been long recognized that one of the most severe mass extinctions in Earth’s history, affecting both the marine and continental biota, occurred at this time [e.g., Newell, 1967; Raup and Sepkoski, 1982; Hallam and Wignall, 1997; McElwain et al., 1999, 2007]. The biotic crisis occurred in the latest Triassic (end-Triassic mass extinction) and is radioisotopically dated at 201.564 ± 0.150 Ma [Blackburn et al., 2013; Davies et al., 2017]. Several hypotheses have been proposed for the trigger of this biotic crisis, including extraterrestrial impact [Olsen et al., 1987, 2002; Morante and Hallam, 1996],

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extensive volcanism (triggering a whole range of paleoenvironmental changes including climate warming, euxinia, ocean acidification, etc.) in the Central Atlantic Magmatic Province (CAMP) [e.g., Marzoli et al., 1999; Schoene et al., 2010; Whiteside et al., 2010; Blackburn et al., 2013; Davies et al., 2017; Percival et al., 2017] and associated climate change [McElwain et al., 1999; van de Schootbrugge et al., 2009; Ruhl et al., 2011], increased photic zone anoxia/euxinia and enhanced ocean stratification [Richoz et al., 2012; Jaraula et al., 2013; Kasprak et al., 2015], or ocean acidification [Hautmann et al., 2008; Hönisch et al., 2012; Greene et al., 2012].

The T–J transition was accompanied by major changes in ocean and atmosphere geochemistry [e.g., Hallam and Wignall, 1997; Pálfy et al., 2001; Cohen and Coe, 2002, 2007; Hesselbo et al., 2002; Pálfy, 2003; Tanner et al., 2004; McElwain and Punyasena, 2007; Hautmann et al., 2008; Kiessling et al., 2009; Kiessling, 2009; Schaller et al., 2012; Bottini et al., 2016], and strontium isotope data suggest a temporary reversal of the long-term decrease in
marine \(^{87}\text{Sr}/^{86}\text{Sr}\) values around the T–J transition [Veizer et al., 1999]. A first indication for a negative carbon isotope excursion (CIE) at the TJB was reported from the Kendelbach section in the Northern Calcareous Alps (NCA) (Austria; Fig. 10.1), but the authors regarded this as likely diagenetically induced [Hallam and Goodfellow, 1990]. McRoberts et al. [1997] showed a negative \(^{13}\text{C}\) peak on bulk rock carbonates originating from the Lorüns section of the NCA; however, this was based only on a single sample that yielded a light value. McElwain et al. [1999] generated two very light \(^{13}\text{C}\) values in organic material from low stratigraphic resolution bulk rock samples from the terrestrial Ashtarkkof succession in Greenland (Fig. 10.1). Combined, these results were suggestive of a potential global carbon cycle perturbation at the T–J transition, and they provided the basis for researchers, including the working group from IGCP project 458 [see Hesselbo et al., 2007], to evaluate the evolution of the global carbon cycle at this time.

Here we review and evaluate temporal fluctuations of isotope ratios and elements in geomatериалs across the TJB, contributing to the development of a comprehensive cheemostratigraphy that can be applied for transcontinental stratigraphic correlation and understanding the processes causing environmental and climatic perturbations that ultimately led to biotic change and mass extinction.

### 10.2. THE END-TRIASSIC MASS EXTINCTION AND POTENTIAL CAUSES

The end-Triassic biotic crisis is one of the big five mass extinction events in Earth’s history [Raup and Sepkoski, 1982]. The taxa affected included ammonoids, scleractinian corals, and conodonts in the oceans [e.g., Guex et al., 2004; Alroy, 2010] and megaflora, sporomorphs, and vertebrates on land [e.g., Benton, 1995; McElwain et al., 1999, 2009; van de Schootbrugge et al., 2009]. The temporal pattern of extinctions and origins through this event and their causes are controversial, and different scenarios have been suggested.

An early hypothesis for what might have caused the end-Triassic mass extinction is a potential impact of a celestial body. Such an event was proposed for the end-Cretaceous mass extinction after finding an iridium concentration in the event beds at Stevns Klint (Denmark) and Gubbio (Italy), reaching highest values of 3 ng/g at the latter locality [Alvarez et al., 1980]. Iridium is rare in crustal rocks (<0.1 ng/g), but relatively enriched in certain meteorite types (>1 μg/g) [Ehmann et al., 1970; Crocket and Teruta, 1977; Crocket, 1979]. An enrichment of Ir was also identified in the Grenzmergel of the TJB section at Kendelbach (Austria) and St Audrie’s Bay (United Kingdom) [Orth, 1989; McLaren and Goodfellow, 1990], but with much lower concentrations compared to that found in Cretaceous-Paleogene (K–Pg) boundary sections, and such small enrichment can also be explained by only invoking volcanic activity [McCartney et al., 1990]. A clear spike of Ir with maximum values of 0.285 ng/g identified by Olsen et al. [2002] was found in a white clay layer between typical Triassic and typical Hettangian pollen and spore assemblages in the Jacksonwald syncline section of the Newark Basin. Even this Ir peak is much smaller compared to those of the K–Pg boundary successions [Alvarez et al., 1980], but it is larger than expected from typical crustal concentration. No clear evidence of other indicators for an extraterrestrial impact, such as impact glass (microtektites, tektites), Ni-rich spinels, micro-spherules, or micro-diamonds, has been found so far at the TJB [e.g., Tanner et al., 2004]. Reports of quartz with planar deformation features from the Grenzmergel at the Kendelbach T–J section in Austria [Badijkov et al., 1987] are probably not impact indicators [Hallam and Wignall, 1997], but rather metamorphic features [Mossman et al., 1998]. Furthermore, the Manicouagan impact structure of Quebec, with ~100 km diameter that represents one of the largest impacts known in the Phanerozoic [Grieve, 1998] and originally favored by Olsen et al. [1987] as that responsible for the end-Triassic mass extinction, is more than 10 million years older than the TJB [Hodych and Dunning, 1992]. Other Triassic craters [for review see Tanner et al., 2004], such as the 80 km Puchezh-Katunki structure in Russia [Ivanov, 1992], are also not of the same age as the end-Triassic mass extinction [Pálfy, 2004; Tanner et al., 2004]. In addition, an \(^{187}\text{Os}/^{188}\text{Os}\) decrease that might be used to identify impact-related strata [Sato et al., 2013] is identified already in Late Triassic Norian strata, therefore pre-dating the TJB.

A more accepted scenario for the cause of the end-Triassic biotic crisis is extensive volcanism. It is well documented that the supercontinent Pangea, dominating the late Paleozoic and Triassic paleogeography for at least 100 million years, began to break up around the T–J transition, substantiated by extensive volcanism in the CAMP, which coincided with the opening of the central Atlantic Ocean [Marzoli et al., 1999; Schlische et al., 2003]. This province had an extent of about 1.1 × 10^7 km^2 including parts of North and South America, northeast Africa, and southwest Europe (Fig. 10.1; e.g., Wignall, 2001a; McHone, 2003; Blackburn et al., 2013; Merle et al., 2014; Pálfy and Kocsis, 2014). CAMP comprises about three million cubic kilometers of basaltic lava, including continental flood basalts that flowed into the large rift basins, and mafic dikes and sills intruded into sedimentary deposits [Marzoli et al., 1999, 2018; McHone, 2003; Saunders, 2005; Davies et al., 2017]. Some CAMP basalt flows, volcanic ashes [Schoene et al., 2010; Blackburn et al., 2013; see also Tanner et al., 2004], or sills
Davies et al., 2017] have been directly related to the end-Triassic mass extinction. It has been suggested that CAMP activity documented by seismites occurring worldwide and by mafic rocks in Morocco started already before the end-Triassic biotic event [Dal Corso et al., 2014; Lindström et al., 2015].

Certainly, vast amounts of CO₂ were exhaled by CAMP volcanism, and, in addition, CH₄ was likely generated by thermal metamorphism of organic-rich sediments [McElwain et al., 2005; Svensen et al., 2007; Korte et al., 2009]. This CO₂ injection has been suggested to be responsible for an abrupt global warming [e.g., Blackburn et al., 2013]. At TJB sections of the Western Carpathians, a sudden cessation of carbonate deposition of the Fatra Formation is evident at the “boundary shale” facies of the Kopieniec Formation, with an inferred increase of riverine influence. This sudden shift has been taken to indicate a sudden climatic change at the erosional TJB [Michalk et al., 2007], potentially induced by an enhanced hydrological cycle upon rising temperatures. However, strong short-term fluctuations superimposed on an overall eustatic sea-level fall have been suggested based on North American and European successions, possibly linked to glacial eustasy and climatic cooling brought about by sulfur degassing in an early stage of the CAMP [Guex et al., 2004, 2012; Schoene et al., 2010]. However, the finding of photic zone anoxia/euxinia resulting from enhanced ocean stratification [Kasprak et al., 2015], as well as the marine biodiversity drop [van de Schootbrugge et al., 2009], has rather been attributed to warming global climates.

10.3. INTENSELY STUDIED TRIASSIC-JURASSIC BOUNDARY SUCCESIONS

Geochemical and chemostratigraphic studies of marine and continental TJB successions have been conducted in recent years, including on some previously proposed GSSPs for the base of the Jurassic. Here we present a synoptic stratigraphic framework and discuss some of the outstanding problems (see Fig. 10.1).

10.3.1. Alpine Sections Including the GSSP Kuhjoch Succession (Austria)

The T–J transition has been extensively studied in paleo-marine shelf settings along the passive margin of the northwestern Tethys Ocean. TJB successions in the NCA in particular are stratigraphically expanded and highly fossiliferous and mainly studied in the Eiberg Basin, Austria, a Rhaetian intraplatform depression, which extends for over 200 km from the Salzkammergut in the east to the Lahnewiesgraben valley (Bavaria) in the west [Hillebrandt et al., 2013]. The intraplatform Eiberg Basin was bordered to the southeast by a broad Rhaetian carbonate platform (Dachstein platform), locally with fringing reefs, and to the southeast by an outer shelf (Hallstatt basin), transitional to the Tethys Ocean [Hillebrandt et al., 2013]. Another partly terrestrial influe- enced carbonate platform, represented by the Oberrhaet Limestone, is located to the north. Intraplatform depressions within the Oberrhaet Limestone also accumulated sedimentary successions across the T–J transition, similar to those in the Eiberg Basin.

The Rhaetian Kössen Formation was deposited over the Hauptdolomit and is composed of limestone and argillaceous bioclastic rocks formed in subtidal systems. The sedimentary facies of the Rhaetian Kössen Formation change around the middle to upper Rhaetian boundary (base of Choristoceras marshi Zone), with the appearance of a basinal facies (Eiberg Member), following on from the shallow-water sequences of the Hochalm Member [Golebiowski, 1989]. The subsiding Eiberg Basin reached 150–200 m water depth in the late Rhaetian [Golebiowski, 1989; Krystyn et al., 2005; Mette et al., 2012; Hillebrandt et al., 2013; and references therein]. Marine conditions therefore prevailed across the T–J transition. However, a distinct and abrupt lithological change from basinal carbonates of the Eiberg Member to marls and clays of the Tiefengraben Member (lower Kendlbach Formation) occurred, possibly in response to a sea-level drop. This siliciclastic stratigraphic unit has also been suggested to coincide with the onset of extrusive CAMP emplacement, partly based on elemental geochemistry and mineralogy for the bituminous topmost layer of the Kössen Formation [Pálfy and Zajzon, 2012]. Observed lithological change at this time may alternatively therefore also reflect changing climatic and environmental conditions, leading to increased precipitation and weathering, and the associated enhanced supply of siliciclastic sediments into the Eiberg Basin.

The lithological transition at the top of the Kössen Formation is marked by the development of a thin (1–5 cm) bituminous layer, with total organic carbon (TOC) values up to 10% [Ruhl et al., 2009]. The overlying reddish gray and faintly laminated mudstones of the Schattwald beds are marked by low TOC and low carbonate concentrations and gradually transition upward into grayish brown marls for the remainder of the Tiefengraben Member [Ruhl et al., 2009]. The Tiefengraben Member (and Kendlbach Formation) were succeeded by Lower Jurassic (upper Hettangian to Sinemurian) carbonate strata of the Adnet Formation, which were deposited under increasing water depths and greater pelagic influence [Böhm et al., 1999].

All sections within the Eiberg Basin show similar sedimentary records across the TJB, with only minor variations in carbonate and clay content depending on more proximal versus more distal positions. The Karwendel syncline...
exposures provide some of the thickest and most expanded and complete marine TJB successions worldwide, which is one of the reasons the Kuhjoch locality was selected as GSSP for the base of the Jurassic system.

The abrupt lithological change from the predominantly carbonate Kössen Formation to the marly sediments of the lower Tiefengraben Mb (locally known as the “Grenzmergel,” which includes the reddish gray colored Schattwald beds) was for long considered to represent the TJB [Golebiowski, 1990; Hallam and Goodfellow, 1990], as it is marked by the disappearance of typical Triassic ammonoid and conodont fossils. Recent studies, however, showed that the lowest meters of the Tiefengraben Member still yield Triassic microflora and nannoflora [Kuerschner et al., 2007].

Newly described psilocreratids in the TJB successions of the Eiberg Basin, including Psiloceras spelae tirolicum, stratigraphically precede the well-known earliest Psiloceras of England (Psiloceras erugatum, Psiloceras planorbis) and the Alps (Psiloceras calliphylum) [Hillebrandt et al., 2013]. The first occurrence of P. spelae tirolicum was selected as marker for the base of the Jurassic system [Hillebrandt et al., 2013]. Within the GSSP succession, P. calliphylum is correlated with the earliest Psiloceras in England [Page, 2003; Bloos, 2004; Hillebrandt and Krystyn, 2009; Page et al., 2010].

The GSSP for the base of the Jurassic and several other TJB sections of the Eiberg Basin were studied in detail for microfossil and macrofossil occurrences [McRoberts et al., 1997; Kuerschner et al., 2007; Bonis et al., 2009a, 2009b; Hillebrandt and Krystyn, 2009; Bonis and Kürschner, 2012; Hillebrandt et al., 2013; and references therein]. Notably, the first occurrence of terrestrial biological markers (pollen/spores) stratigraphically close to the base of the Jurassic allows also for correlation to continental sedimentary sequences of this age [Kuerschner et al., 2007; Bonis et al., 2009a; Bonis and Kürschner, 2012].

Carbon isotope analyses of bulk sedimentary organic matter as well as individual higher plant-derived leaf-wax n-alkanes show a pronounced δ13C negative excursion of ~6 and ~8‰, respectively, in the studied sections of the Eiberg Basin [Ruhl et al., 2009, 2011]. This negative CIE occurs at the very base of the Tiefengraben Member, directly coinciding with the end-Triassic mass extinction, and it is marked by a TOC-rich bituminous black shale at its onset [Bonis et al., 2009b]. Further to the west, in the Lechtal nappe of the Bajuvaric nappe group, in the western NCA (western Austria), the T–J transition is recorded within broad carbonate platform sedimentary sequences [Felber et al., 2015]. Also here, the upper Triassic Kössen Formation was deposited on top of the Hauptdolomit (of which the upper part is informally known as “Plattenkalk”) and followed by progradational siliciclastic sediments [Berra et al., 2010]. The transitional Schattwald beds including the TJB are here followed by the Hettangian “Lorüns oolite” [Felber et al., 2015].

Specific clay minerals (low- to medium-charged smectite and Mg-vermiculite), altered but euhedral mafic minerals, and elevated heavy REE concentrations in the lower Tiefengraben Member of the Eiberg Basin were suggested to have been associated with the onset of CAMP volcanism and the associated atmospheric dispersal of volcanic ash and weathering processes [Pálfy and Zajzon, 2012]. The combined biostratigraphic and chemostratigraphic framework of the TJB sections in the Austrian Eiberg Basin provides an excellent framework for the stratigraphical correlation to marine and continental T–J sedimentary sequences elsewhere; for the study of oceanographic, climatic, and environmental changes at that time; and for understanding their temporal and potentially causative link to CAMP volcanism.

10.3.2. Bristol Channel Basin at St Audrie’s Bay
(United Kingdom)

The T–J sedimentary succession in the Bristol Channel Basin has long been studied for microfossil and macrofossil content, and the outcrops at St Audrie’s Bay were proposed as GSSP for the base of the Hettangian stage (base of the Jurassic system) [Warrington et al., 1994, 2008, and references therein]. The recognition of a negative excursion in δ13CTOC coinciding with the end-Triassic mass extinction event in this sedimentary record sparked the study of the potential (temporal) link between CAMP volcanism, global carbon cycle change, and biotic response at this time interval [Hesselbo et al., 2002]. The ~5‰ negative CIE, informally called the “initial negative CIE,” at the end-Triassic mass extinction level is followed by a return to more positive values and a subsequent long-term shift to yet more negative values (the “main CIE”) [Hesselbo et al., 2002]. This evolution in δ13CTOC from the mass extinction level upward is also mirrored in the δ13CFossil calcite record from the same and other nearby sections [Korte et al., 2009]. Facies analysis in the Bristol Channel Basin sedimentary succession suggests a sea-level lowstand stratigraphically just preceding the end-Triassic mass extinction and “initial” negative CIE [Hesselbo et al., 2004].

Paleomagnetic, chemostratigraphic, biostratigraphic, and cyclostratigraphic analyses of the sedimentary record at St Audrie’s Bay have provided a detailed stratigraphic framework for correlation to marine and continental sedimentary records in other regions worldwide and to the CAMP magmatic records of Morocco and North America [Hounslow et al., 2004; Warrington et al., 2008; Bonis et al., 2010; Deenen et al., 2010; Ruhl et al., 2010; Bonis and Kürschner, 2012; Mander et al., 2013; Hüsing et al., 2014; Xu et al., 2017].
10.3.3. Csóvár Section in the Transdanubian Range Unit (Hungary)

The sections near the village of Csóvár are located in north-central Hungary (Fig. 10.1), near the northeastern end of the Transdanubian Range Unit segment of the Tethyan shelf [Haas et al., 2010]. The T–J transition is recorded in carbonate rocks deposited in open marine, basinal to toe-of-slope settings of the Norian to Sinemurian Csóvár Limestone Formation [Pálfy and Haas, 2012] deposited in a periplatform basin. Dark gray, bituminous calcarenite (containing shallow-water fossils and clasts supplied from the adjacent platform) and marl beds are predominant in the lower part, and well-bedded, pale yellow to pale brown micritic limestone and cherty limestone composes the upper part [Haas et al., 1997; Haas and Tardy-Filácz, 2004] (note that Kozur [1993] suggested that the upper unit should be distinguished as a separate formation, the Várhegy Cherty Limestone Formation). Two different sections of partly overlapping age are exposed on both sides of the Pokol-völgy (Hell Valley) northwest of the Csóvár village: the Pokol-völgy quarry and the Vár-hegy (Castle Hill) section (see map, e.g., in Pálfy et al., 2007). An uppermost Rhaetian marker bed containing abundant lithoclasts and platform-derived bioclasts helps to establish lithologic correlation between the two sections. Syn-sedimentary slump structures occur at several levels in the lowermost Jurassic and potentially complicate interpretation of the isotope curve.

Ammonoid biostratigraphy provides broad constraints for drawing the TJB between occurrences of Rhaetian Choristoceras sp. and an early Hettangian psiloceratid, with the intervening 17 m only yielding an ex situ specimen of Nevadaphyllites [Pálfy and Dosztály, 2000; Pálfy et al., 2007]. Radiolarians indicating the Globolaxtorum tozeri and Canoptum merum zones (latest Rhaetian and earliest Hettangian, respectively) also bracket a ~25 m barren interval [Pálfy et al., 2007]. The staggered disappearance of conodonts falls into this interval, subdivided into the Misikella ultima and Neohindeodella zones [Pálfy et al., 2001, 2007; Korte and Kozur, 2011]. However, as Kozur [1993] already noted, the Csóvár section may be unique in preserving the last and youngest conodont taxa, with rare survivors reaching into the earliest Jurassic. On palynological grounds, a synchronous marine and terrestrial TJB event is suggested by correlated spikes in the abundance of prasinophyte algae and fern spores [Götz et al., 2009].

The Vár-hegy section provided one of the first detailed C isotopic curves across the TJB and the first one where a T–J transition negative anomaly was simultaneously documented in both δ13Ccarb and δ13Corg, with a magnitude of −3.5 and −2‰, respectively, over a stratigraphic interval of ~2 m within the available biostratigraphic brackets of the system boundary [Pálfy et al., 2001]. Despite some noise in the δ13Ccarb data attributed to minor diagenetic overprint, this anomaly stands out and is confirmed by the δ13Corg data, lending support to its primary character. Carbon isotope ratios are not significantly correlated with either TOC or hydrogen index values. Another comparably minor negative anomaly is observed somewhat lower in the upper Rhaetian. A subsequent high-resolution study extended the δ13Ccarb curve for an additional 20 m into the middle Hettangian and focused on the TJB negative anomaly [Pálfy et al., 2007]. Here the TJB anomaly appears to contain a series of short-term oscillations, whereas no clear trend was found after the curve leveled off returning to the pre-exursion values. A δ13Ccarb curve was also obtained from a 16 m thick upper Rhaetian part of the Pokol Valley quarry section by Korte and Kozur [2011]. Although a <1‰ negative excursion followed by a stepped ~2‰ positive shift is recognized, its correlation to the nearby Vár-hegy section is not straightforward. Additional geochemical investigations of Rhaetian conodonts from the Pokol Valley quarry are required to help refine the δ Sr reference curve focused on the TJB [Korte, 1999; Korte et al., 2003]. Although the Csóvár sections yield important data for our understanding of the carbon cycle perturbation at the TJB, they pose a challenge regarding the curves obtained to date and the correlation of them with other sections. The main negative anomaly in the Vár-hegy section is best equated with the initial CIE, and a precursor anomaly also appears to be recorded. However, the main negative anomaly reported from other sections remains unproven here.

10.3.4. Kennecott Point in Haida Gwaii (Queen Charlotte Islands, Canada)

Studies of an expanded TJB section at Kennecott Point yielded δ13C curves with one of the earliest recognized anomalies across the TJB [Ward et al., 2001]. Haida Gwaii (also known as the Queen Charlotte Islands), a Pacific archipelago in British Columbia, forms part of the accreted terrane of Wrangellia and preserves a relatively continuous and unmetamorphosed Middle Triassic to Middle Jurassic sedimentary succession [Lewis et al., 1991]. Kennecott Point is located on the northwestern shore of Graham Island, with excellent exposures on a wave-cut intertidal platform. Siliciclastic sediments (thin-bedded, organic-rich shale and siltstone with fine- to medium-grained sandstone interbeds) of the Rhaetian to Sinemurian Sandilands Formation were deposited in deep marine outer shelf to basin systems [Desrochers and Orchard, 1991].

The TJB has been defined by integrated ammonoid, radiolarian, and conodont biostratigraphy [Tipper et al., 1994]. Radiolarians occur abundantly throughout the
section, enabling a highly resolved biostratigraphy. A marked faunal turnover leads to recognition of the boundary between the *Globolaxatorum tozeri* and *Canoptum merum* zones, which is also equated to the system boundary [Carter and Hori, 2005; Longridge et al., 2007]. The Norian-Rhaetian boundary is approximated by the base of the *Parvicingulum moniliformis* radiolarian zone [Carter, 1993] or the last appearance of the distinctive bivalve *Monotis* [Williford et al., 2004]. Sparse latest Triassic ammonoids include *Choristoceras rhaeticum* and *Choristoceras nobile*, assigned to the North American *Choristoceras crickmayi* Zone [Tozer, 1994; Ward et al., 2001], whereas the *Choristoceras minitus* Zone, the second lowest Jurassic ammonoid zone, is documented by the index species *Choristoceras minitus*, appearing ~8 m higher than the earliest Jurassic radiolarians, and followed by *Psiloceras* ex gr. *tilmanni*, indicative of the next higher *Psiloceras pacificum* Zone of the lower Hettangian [Longridge et al., 2007]. The Rhaetian zonal index conodont *Misikella posthernsteini* was found in a single sample in the uppermost Triassic part of the section, and the highest conodont occurrence is recorded close to the radiolarian-defined TJB [Tipper et al., 1994]. Significantly, there is no change in lithology at the TJB in the Kennecott Point section.

Ward et al. [2001] obtained a $\delta^{13}$C$_{org}$ curve from a ~120 m thick part of the section, from the uppermost Norian through the lowermost Hettangian. This study documented a significant, $\sim 2\%o$ excursion over <5 m of strata directly at the TJB, following the Rhaetian with no obvious trend in the C isotopic evolution. No significant correlation was found between TOC and $\delta^{13}$C$_{org}$ values, lending support to the interpretation as a primary signal. Subsequent work at higher sampling resolution confirmed the presence of the TJB negative anomaly and described it as a series of short-term oscillations with up to six local minima [Ward et al., 2004]. On the other hand, isotopic values measured on bulk carbonate from limestone concretions and interbeds appeared diagenetically overprinted and were not considered for further interpretation. Williford et al. [2007] extended the isotopic analyses to the higher, Lower Jurassic part of the section, doubling the thickness of the sampled stratigraphic section to 250 m, through the entire Hettangian and well into the Sinemurian. However, as faulting is known to cause tectonic repetitions at Kennecott Point [Pálfy et al., 1994; Longridge et al., 2008], caution is needed as the measured and sampled section may include some overlooked tectonically duplicated parts. The key feature of the extended $\delta^{13}$C$_{org}$ curve is the presence of a pronounced 5%o positive excursion in the Hettangian, closely following a transient return from the TJB negative spike to pre-exursion values. Although correlation of this Lower Jurassic isotope curve together with ammonoid and radiolarian biostratigraphic data is not tightly constrained, the positive anomaly appears early to mid-Hettangian in age. Post-exursion $\delta^{13}$C$_{org}$ values in the upper part of the Hettangian and Sinemurian section are $\sim 1\%o$ more negative than the long-term Rhaetian average values and the values obtained for the transient return after the TJB negative anomaly.

Sulfur isotopic analyses of a suite of samples revealed a major positive $\delta^{34}$S anomaly coincident with the Hettangian positive $\delta^{13}$C$_{org}$ excursion [Williford et al., 2009]. On the other hand, the TJB negative CIE seems to correspond to a significant protracted negative $\delta^{13}$S anomaly, as well as the occurrence of lipid biomarkers suggestive of photic zone euxinia [Kasprak et al., 2015].

In summary, the Kennecott Point section with continuous deep marine sedimentation in the East Pacific realm provides a useful reference to define and understand perturbations of the global carbon, sulfur, and nitrogen cycles, whereas the fossil record is utilized for both biostratigraphic constraints and assessment of biotic changes during mass extinction.

### 10.3.5. New York Canyon Section (Nevada, United States)

One of the best known and most intensively studied TJB sections in North America (and the world) is at Ferguson Hill and Muller Canyon in the New York Canyon area of the Gabbs Valley Range (Mineral County, Nevada), ~170 km southeast of Reno. The significance of this continuous marine TJB section was first recognized by Muller and Ferguson [1936] who subdivided the predominantly dark shale, siltstone, and limestone strata across the system boundary into the Gabbs and Sunrise formations, with a gradational contact between them. Further lithostratigraphic subdivision led to the introduction of members [Taylor et al., 1983], of which, in ascending order, the limestone-dominated Mount Hyatt and siltstone-dominated Muller Canyon members of the Gabbs Formation and the limestone-dominated Ferguson Hill Member of the Sunrise Formation were the subjects of several biostratigraphic and chemostratigraphic studies. Hallam and Wignall [2000] drew attention to the local expression of the extinction and discussed its relation to facies changes. The continuous shallow marine sedimentary succession was deposited in a foreland basin east of the Sonoma allochthon, following Permian-Triassic thrusting related to the Sonoma orogeny at the Cordilleran margin of North America [Dickinson, 2004]. Magnatism related to a later Cretaceous phase of tectonic evolution led to a low-grade metamorphic overprint, hindering magnetostratigraphic studies and destroying palynomorphs [Lucas et al., 2007].

The macrofossil record across the TJB is particularly rich in ammonoids [Guex, 1995] and bivalves [Laws,
1982], which served as the basis for the GSSP candidacy of the section [Guex et al., 1997; Lucas et al., 2007; McRoberts et al., 2007]. Latest Triassic Choristoceras spp. occur up to the lowermost Mount Hyatt Member, below a 7 m thick barren interval, followed by the first occurrence of Psiloceras spelae and P. tilmanni, regarded as the oldest Jurassic ammonoid species and zonal indices of the lowermost Hettangian ammonoid biozone. The first occurrence of pectinid bivalve Aegerchlamys boellingi is also of stratigraphic significance near the system boundary, immediately below the first psiloceratid ammonoids [McRoberts et al., 2007]. The age of latest Triassic strata is also well supported by conodont and radiolarian biostratigraphy [Orchard et al., 2007]. A high-precision U-Pb zircon age of 201.33 ± 0.13 Ma provides a numeric tie point to calibrate other stratigraphic schemes and the geological time scale and to correlate with other radioisotopically dated sections [Schoene et al., 2010].

The first δ13Corg curve from the New York Canyon area was produced by Guex et al. [2003, 2004], documenting two negative excursions of similar ~2‰ amplitude, the first one near the last occurrence of C. crickmayi and the upper one between the first occurrence of P. spelae, P. tilmanni, and P. pacificum. Despite some scatter in the data, a return to pre-exursion values is observed between the anomalies. Ward et al. [2007] reported a new set of δ13Corg data from an independently collected suite of samples. Although it confirmed the presence of two negative anomalies in a curve with less scatter, it also led to controversies regarding the position of the negative anomalies with regard to lithostratigraphy and biostratigraphic markers, recording a positive peak in the earliest Jurassic and an apparent offset in values of isotopic ratios compared to those reported in Guex et al. [2004]. Guex et al. [2009] argued that part of these discrepancies is explained by differing views on the definition of lithostratigraphic units and the overlooking of a fault by the other authors. Comparison of the stratigraphic positions of both negative anomalies in δ13Corg therefore remains ambiguous. Repeated measurements on five samples yielded values >0.3‰ different from the originally reported ones. A third independent sampling was carried out, and new measurements were reported by Thibodeau et al. [2016], with the resultant curve in good agreement with a corrected version of that of Ward et al. [2007]. These data were obtained for a geochemical study which also documented elevated Hg concentrations, with a peak at the termination of the first negative C isotope anomaly, supporting the inference of a volcanic trigger for the environmental changes [Thibodeau et al., 2016]. An additional 20 m of section was sampled up to the Hettangian-Sinemurian boundary, and the extended curve features a prominent 5‰ positive anomaly in the upper Hettangian [Bartolini et al., 2012]. Thus, the New York Canyon section also contributes to our understanding of the Hettangian carbon isotope record, where correlation of positive anomalies remains controversial.

In summary, geochemical studies from the New York Canyon area (i) span the T-J transition and the entire Hettangian; (ii) benefited from good ammonoid biostratigraphical control; (iii) are important in establishing the succession of two separate negative carbon isotope anomalies, followed by a prominent positive anomaly; and (iv) are unparalleled in being based on three independent sets of samples and measurements by different research teams.

10.3.6. Astartekløft (East Greenland)

Astartekloft is the best studied of several localities in the Hurry Inlet (Scoresby Sund) area of east Greenland, part of the Jameson Land Basin, which have been of particular interest for their paleobotanical contents. Within the Jameson Land Basin, the TJB occurs in the fluviolacustrine strata of the Kap Stewart Group [Surlyk, 2003], which range from marginal fluvial sandstone to basin-center shale. The Hurry Inlet localities show a marginal fluvial succession comprising channels filled with coarse sand, commonly multistory, with thinner overbank deposits that include plant-bearing crevasse splays [Dam and Surlyk, 1992; McElwain et al., 2007]. Initial thorough stratigraphic work by Harris [1937] documented in the Hurry Inlet localities a significant turnover of florals from the supposed Triassic Lepidopteris flora to the Jurassic Thamnophyta flora. Harris’ work was subsequently expanded upon by McElwain et al. [2007, 2009] who amassed large collections of plant macrofossils from Harris’s plant beds at Astartekloft. Several studies have built upon this paleobotanical framework to suggest significant changes in plant ecosystems across the TJB in response to large igneous province (LIP) forcing [e.g., Belcher et al., 2010; Bacon et al., 2013; Mander et al., 2013; Steinthorsdottir et al., 2018], and the Astartekloft fossil plant (stomatal density/index) record has been the principal basis for reconstruction of atmospheric CO2 changes across TJB [McElwain et al., 1999; Steinthorsdottir et al., 2011, 2012]. Carbon isotope chemostratigraphic correlation was used by Hesselbo et al. [2002] for correlation to St Audrie’s Bay in southern England. The carbon isotope stratigraphy of Hesselbo et al. [2002], based on macrofossil wood, was corroborated by analysis of specifically identified leaf cuticles [Bacon et al., 2011]. On the basis of detailed palynological study, Mander et al. [2013] suggested a revised correlation to St Audrie’s Bay in which strata equivalent to the “initial” CIE are missing or undetected at Astartekloft.
10.4. CHEMOSTRATIGRAPHY

10.4.1. Carbon Isotope Stratigraphy and Total Organic Carbon (TOC) Variation

Following the pioneering work by McRoberts et al. [1997] and McElwain et al. [1999], many studies investigated the carbon isotope trend across the TJB. Negative CIEs were confirmed from other marine [e.g., Pálfy et al., 2001; Ward et al., 2001; Hesselbo et al., 2002; Guex et al., 2004; Galli et al., 2005; Kerschner et al., 2007; McRoberts et al., 2007; Williford et al., 2007; Ruhl et al., 2009] and terrestrial [Hesselbo et al., 2002; Steinhorsdottir et al., 2011; Pieńkowski et al., 2012] TJB sections and have been related to the end-Triassic mass extinction. In the western NCA, the isotope excursion is marked by a ~3‰ negative shift in bulk carbonate δ13C [Felber et al., 2015], which is similar to that recorded at a northern Italian section (3–5‰ negative CIE), the Budva Basin in Montenegro (1–2‰ negative CIE), and the United Arab Emirates (~5‰ negative CIE) [Galli et al., 2005; Crne et al., 2011; Al-Suwaidi et al., 2016].

The negative CIE in the Eiberg Basin is followed by a return to pre-exursion values throughout the Schattwald beds (lower Tiefengraben Member) and the subsequent gray marls of the upper Tiefengraben Member (Fig. 10.2), with 1–2‰ more negative δ13C values broadly coinciding with the first occurrence of *Psiloceras spelaee* tirolicum at the base of the Jurassic [Ruhl et al., 2009]. A shift to the continuously lighter δ13C values of the Hettangian stage, as, for example, observed in the marine Bristol Channel Basin [Hesselbo et al., 2002; Korte et al., 2009; Ruhl et al., 2010], the Danish Basin [Lindström et al., 2012], in the New York Canyon section of Nevada [Bartolini et al., 2012], and the continental Newark and Hartford basins [Whiteside et al., 2010], occurs in the Eiberg Basin broadly at the level of the first occurrence of the *Psiloceras cf. pacificum* [Hillebrandt et al., 2013].

Several studies have suggested a minor carbon cycle perturbation and associated negative CIE to precede the end-Triassic mass extinction (sometimes referred to as the precursor CIE) [Deenen et al., 2011; Ruhl & Kürschner, 2011; Lindström et al., 2012; Dal Corso et al., 2014; Davies et al., 2017]. The Upper Triassic sequences in the Eiberg Basin have only been studied extensively in the Eiberg quarry [Korte et al., 2017]. Based on this succession, an even earlier carbon cycle perturbation was suggested to have occurred in the late Rhaetian (the late Rhaetian CIE; Mette et al., 2012).

Observed trends in δ13C records led to the suggestion of a characteristic δ13C geometry for the TJB interval including a short-term “initial” negative excursion, followed by a longer-lasting positive excursion and the long-lasting “main” negative excursion in marine successions [Hesselbo et al., 2002; Kerschner et al., 2007; McRoberts et al., 2007; Ward et al., 2007; Williford et al., 2007; Korte et al., 2009; Ruhl et al., 2009, 2010, 2011; Korte & Kozur, 2011] and that the end-Triassic mass extinction is coeval with the “initial” negative peak [e.g., Guex et al., 2004; Hesselbo et al., 2007; McRoberts et al., 2007; Ruhl et al., 2009; Korte & Kozur, 2011]. This trend in δ13C has also been identified in some terrestrial successions [Deenen et al., 2010; Whiteside et al., 2010; Dal Corso et al., 2014], reinforcing that it is likely global in nature and that it reflects changes in δ13C values of the global ocean-atmosphere system. Additional potential excursions predating that across the TJB have been also discovered [Cleveland et al., 2008; Ruhl & Kürschner, 2011; Schaller et al., 2011; Mette et al., 2012; Korte et al., 2017]. The clear major (“main”) negative TJB excursion, however, has not be identified in all marine sections [e.g., see Galli et al., 2007; Pálfy et al., 2007; van de Schootbrugge et al., 2008; Götz et al., 2009]. This has led some others to suggest that the amplitude, shape, duration, and even the stratigraphic position of the “initial” δ13C minimum are different between sections and therefore do not represent a clear chemostratigraphic marker [Lindström et al., 2017]. This inference, however, largely stems from limited data on the abundance of scarce pollen and spore species and sometimes neglects all other available stratigraphic markers, such as ammonites and the first and last occurrences of specific polyomorphs. Detailed stratigraphic correlation between individual TJB successions can be achieved through an integrated biostratigraphic and chemostratigraphic framework (Figure 10.2b).

The release of isotopically light carbon, such as CO₂ from volcanic degassing, thermal metamorphism of organic-rich sediments, and/or biogenic methane likely resulted in a δ13C negative shift in global exogenic carbon reservoirs. Because of this and the relatively short (10⁴–10⁵ yr) residence time of carbon in the ocean-atmosphere-biosphere, peaks and troughs in carbon isotope records should represent useful markers for transcontinental (and marine-terrestrial) correlations.

Only a limited amount of compound-specific δ13C data has been published to date, only on leaf-wax-derived long-chain n-alkanes [Whiteside et al., 2010; Ruhl et al., 2011].

The observed changes in these organic geochemical compounds are suggested to reflect large changes in T–J atmospheric δ13C, and they do directly correspond to changes in bulk organic δ13C (δ13C_TOC and δ13C_wood) in the same sedimentary successions [Whiteside et al., 2010; Ruhl et al., 2011]. Further similar work spanning the Upper Triassic and Lower Jurassic is necessary, but it has been limited by, for example, thermal maturity, low sedimentary TOC values, and poor organic sedimentary preservation in available successions.
Figure 10.2 (a) Comparison of Triassic-Jurassic $\delta^{13}$C$_{carb}$, $\delta^{13}$C$_{TOC}$, and $\delta^{13}$C$_{wood}$ data from geographically distributed marine and terrestrial sections, with Hg/TOC ratios and inferred atmospheric $p$CO$_2$. Atmospheric $p$CO$_2$ estimates are based on stomatal density analyses (Astartekløft, Greenland) and $\delta^{13}$C values of pedogenic carbonate sequences (Newark and Hartford basins). Note that this compilation is not exhaustive and that many more Triassic-Jurassic boundary sections have been studied. Data in (a) are from: Newark and Hartford basins, USA: kent et al. [2017] and references therein, Schaller et al. [2011, 2012, 2015]; St Audrie’s Bay, UK: Hesselbo et al. [2002], Korte et al. [2009], Ruhl et al. [2010], Hüsing et al. [2014], Xu et al. [2017]; Kuhjoch, Austria: Ruhl et al. [2009], Hillebrandt et al. [2013], Percival et al. [2017]; New York Canyon, Nevada, USA: Bartolini et al. [2012], Thibodeau et al. [2016]; Astartekløft, Greenland: Hesselbo et al. [2002], Steinthorsdottir et al. [2011], Percival et al. [2017]; Csóvár, Hungary: Pályi et al. [2001]; Canj, Montenegro: Crne et al. [2011]. (b) Highly resolved Triassic-Jurassic boundary integrated stratigraphic framework based on key European successions studied for $\delta^{13}$C$_{TOC}$ palynostratigraphy, and ammonite biostratigraphy. Data in (b) are from: Stenlille 1/2: Lindström et al. [2017] and references therein; St Audrie’s Bay, UK: Hesselbo et al. [2002] and references therein, Hounslow et al. [2004], Bonis and Kürschner [2012]; Tiefengraben, Austria: Kuerschner et al. [2007], Hillebrandt et al. [2013]; Kuhjoch (integrated Kuhjoch East & West sections), Austria: Bonis et al. [2009a], Ruhl et al. [2009], Hillebrandt et al. [2013]; Hochalplgraben, Austria: Bonis et al. [2009b], Ruhl et al. [2009], Hillebrandt et al. [2013].
10.4.2. Mercury Chemostratigraphy

The concentration of Hg in the ocean-atmosphere system is largely controlled by emission as a trace volcanic gas from continental volcanoes or mid-ocean spreading ridge systems [Pyle & Mather, 2003; Bowman et al., 2015]. Gaseous elemental Hg has a typical residence time of 0.5–2 years, allowing for global atmospheric dispersal before eventual drawdown into marine and continental sediments [Blum et al., 2014]. Importantly, Hg is typically drawn down into sediments bound with organic matter [Benoit et al., 2001; Outridge et al., 2007], although chemical binding to sulfides and clays may also be of some importance [Benoit et al., 1999; Niessen et al., 2003; Kongchum et al., 2011; Bergquist, 2017]. Sedimentary Hg concentrations are therefore typically normalized against TOC content of the sediment [Percival et al., 2017]. The timing of activity and impact of several Phanerozoic LIPs and their potential relationships to global change (e.g., such as oceanic anoxic events (OAEs)) and mass extinction events, including the end-Permian, early Toarcian, mid-Cretaceous (Cenomanian-Turonian) OAE2, and end-Cretaceous events, were previously already studied [Sanei et al., 2012; Percival et al., 2015, 2016; Font et al., 2016; Sial et al., 2016; Scaife et al., 2017].

The potential temporal correlation between events at the T–J transition, including the end-Triassic mass extinction, and the emplacement of CAMP was recently extensively studied in marine and continental sedimentary records from both hemispheres and from different paleolatitudes [Thibodeau et al., 2016; Percival et al., 2017]. An initial major pulse in sedimentary Hg concentrations, possibly reflecting the onset of Hg emissions, directly coincides with the end-Triassic mass extinction event [Percival et al., 2017] (Fig. 10.2a). More importantly, however, individual major CAMP basalt flows preserved in the continental basins of North America (the United States and Canada) and North Africa (Morocco) can potentially be temporally (and perhaps causatively) linked to peaks in sedimentary Hg accumulation [Percival et al., 2017].

A recent study suggests that extrusive emplacement of CAMP may have been preceded, by ~100 kyr, by the intrusive emplacement of dike and sill systems, partly in sedimentary basins, which caused an early perturbation stage in the Earth’s climate [Davies et al., 2017]. The analyses of Hg concentration have so far been largely stratigraphically focused at the TJB. Future studies may provide further insights on Hg release, which may also have been generated from thermogenic processes around the sill complexes that intruded sedimentary basins already in the latest Triassic.

10.4.3. Oxygen Isotope Stratigraphy

The oxygen isotopic composition of marine carbonates is generally controlled by seawater temperature [Urey et al., 1951] as well as by seawater δ18O and pH [Zeebe & Wolf-Gladrow, 2001]. The temperature dependence, leading to lighter oxygen isotope ratios in carbonates secreted in warmer water and vice versa, enables the use of δ18O_carb as paleothermometer for ancient oceans. In addition, the carbonate carbon isotope (Section 10.4.1) measurements are performed in parallel and on the same aliquots as those for the oxygen isotopes, allowing a direct comparison, for example, reconstructing temperature (climate) change together with atmosphere-ocean fluctuations of carbon dioxide. Oxygen isotope ratios, especially those from bulk carbonates, are, in comparison to carbon isotopes, more prone to diagenesis [e.g., Veizer, 1983; Marshall, 1992]. A careful evaluation of the data is therefore necessary to interpret seawater temperature changes of the past. On the basis of a dataset from which potentially altered bulk carbonate δ18O data were culled [Pálfy et al., 2001, 2007], an extreme temperature increase of more than 10 °C across the TJB at the Csővár section in Hungary is suggested, and this is quite dramatic for paleolatitudes of about 30°N (Fig. 10.1). A similar bulk carbonate δ18O negative shift is also evidenced in other successions, such as the Doniford section in southwest England [Clémence et al., 2010], suggesting marine climate change with a superregional extent. Pálfy et al. [2001, 2007] pointed out that this severe apparent temperature increase occurred during the period when the initial negative CIE happened (see Fig. 10.2) and during a time span when an injection of isotopically light carbon from volcanic exhalation of CO2 and/or a methane hydrate release occurred, which potentially triggered a global warming [Pálfy et al., 2001; Hesselbo et al., 2002, Ruhl et al., 2011]. This hypothesis of climate warming has also been suggested by model data [e.g., Huynh & Poulsen, 2005] and the findings of shelf-sea photic zone anoxia or even euxinia [Jaraula et al., 2013; Kasprak et al., 2015].

However, other bulk carbonate δ18O datasets of this interval show heavier values (Lorüns section in Austria; McRoberts et al., 1997), or lighter values (Kendelbach section in Austria) have been interpreted as the result of diagenetic alteration [Morante & Hallam, 1996]. For a robust evaluation of a global temperature rise, more proxy data from pristine samples across the TJB are still necessary, and this is in concert with the conclusion by Tanner et al. [2001], evaluating carbon isotopes on pedogenic carbonates.

Oxygen isotope values from low-Mg calcite fossils, such as brachiopods and oysters, represent a robust dataset useful for reconstruction of past seawater temperatures because this material is relatively resistant to diagenesis.
Figure 10.3 Compilation of continental Triassic-Jurassic sequences, which have been analyzed for δ¹³C_TOC and δ¹³C_wood and which record Central Atlantic Magmatic Province (CAMP) basalt emplacement. Data from: Hartford Basin, USA: Schaller et al. [2012], Kent et al. [2017] and references therein; Newark Basin, USA: Schaller et al. [2011], Kent et al. [2017] and references therein; Fundy Basin, Canada: Deenen et al. [2011]; Argana Basin, Morocco: Deenen et al. [2010]; High Atlas Mountains, Morocco: Dal Corso et al. [2014], El Ghilani et al. [2017]. (See insert for color representation of the figure.)
[e.g., Popp et al., 1986; Veizer et al., 1986] and because the alteration degree can be assessed by a multitude of physical and chemical techniques [see Veizer, 1983; Marshall, 1992; Ullmann & Korte, 2015 for reviews]. From different regions in Europe, δ18O (and δ13C) data exist from well-preserved brachiopods of the Late Triassic [Korte et al., 2005, 2017; Mette et al., 2012] and pristine oysters of the earliest Jurassic [Jones, 1992; Korte et al., 2009]. Unfortunately, no continuous dataset over the TJB is available from a single locality. A combination of datasets from different regions (with a lack of data in the latest Triassic and earliest Jurassic [Korte et al., 2009, 2017]) is a relatively poor basis for reconstructing the temperature evolution across the whole time period of interest. However, the dataset from Korte et al. [2009] provides insights about the bottom water temperature changes during the earliest Jurassic in the Bristol Channel Basin, United Kingdom (Fig. 10.4). Relatively cool seawater (assuming a seawater δ18O of −1.2‰ SMOW) temperatures between <7 and 14°C existed for (gray field in the biozone row of Fig 10.4) the period of deposition of the upper Langport Member around the TJB. The temperature increased then distinctly by more than 8°C, reaching values ~12 and 22°C in the first 3 ammonite subzones of the Jurassic. Extreme changes in seawater values ~12 and 22 °C in the first 3 ammonite subzones of the Jurassic. Extreme changes in seawater δ18O (evaporation, melting of continental ice, meteoric water dilution) as a trigger for these severe oxygen isotope fluctuations have been excluded by Korte et al. [2009]. This oxygen isotope signal therefore potentially reflects a true climate signal in which the warming (Fig. 10.4) occurred together with the trend toward to lighter carbon isotope values (main negative CIE of Hesselbo et al. [2002]). The data are compatible with the suggestion that heavy carbon isotope values correspond to periods of lower pCO2 contents and vice versa. However, the detailed comparison between the δ18O and the δ13C curves shows small differences in the positions of the positive peaks (Fig. 10.4). The δ18O decrease (i.e., rise in temperature) in this case began earlier than the δ13C decline and with it the inferred increase in CO2 levels. This suggests that factors other than CO2 may have contributed to the warming in this area, such as opening of the seaway to marine currents derived from warmer, more equatorial waters as also suggested for the Laurasian seaway in the Aalenian [Korte et al., 2015].

We note, however, that the relative heavy oxygen isotope values of nearly +2‰ (temperatures between <7 and 14°C) in the upper Langport Member (gray field in in the biozone row in Fig. 10.4) reflect similarly cool temperatures evidenced for the Aalenian stage in the Hebrides Basin of Scotland [Korte et al., 2015; see also Korte & Hesselbo, 2011; Korte et al., 2017]. These data rather indicate a short-term “cold mode” interval existed around the TJB, at least equivalent to the upper Langport Member at mid-European paleolatitudes, which were south of 40°N (Fig. 10.1). Cool temperatures close to the T–J transition were also identified by sporomorph associations from several successions in central and northwestern Europe and in northeast Greenland [Hubbard & Boulter, 2000]. Although this was later challenged by McElwain et al. [1999], pollen data do suggest initial warming at the end-Triassic mass extinction interval to be followed by a short cooling phase at or directly preceding the TJB and a long-term warming into the Hettangian stage [Bonis & Kürschner, 2012].

These relatively cool temperatures at the TJB (upper Langport Member) occur during the interval of subaerial basaltic eruptions from the CAMP coeval with Hg peaks (Figs. 10.2a and 10.4; Section 10.4.2; see also Fig. 10.3) and suggest volcanism-induced cooling as a plausible explanation [Guex et al., 2004, 2012; Schoene et al., 2010]. It has been proposed that volcanic eruptions have triggered short-term climate cooling events [Lamb, 1970; Kennett & Thunell, 1975; Rampino et al., 1988] when erupted ashes, sulfur dioxide, and hydrogen sulfide (hydrogen sulfide rapidly oxidizes to SO2 in the atmosphere) reach the stratosphere. Ashes fall down rapidly, whereas SO2 reacts with hydroxyl (OH−) and water to form sulfuric acid and generates aerosol clouds which circulate for several years around the planet [Cadle et al., 1976; Pollack et al., 1976; Rampino et al., 1988]. These clouds reduce the transparency for light and, in addition, reflect solar radiation, and these factors cause the troposphere and Earth’s surface to cool down [Devine et al., 1984; Sigurdsson, 1990]. Volcanic cooling events, induced by short-term felsic (high-SiO2) or intermediate volcanic eruptions, are usually of short duration and lasting not longer than two years (e.g., Mt. Pinatubo eruption in 1991; Genin et al., 1995), and it was therefore debated whether volcanically induced cool periods can last over hundreds of years [Bryson & Goodman, 1980; Rampino et al., 1988]. CAMP emplacement, however, represents the most extensive subaerial LIP in Earth’s history [McHone, 2003; Blackburn et al., 2013]. Its emplacement spanned 600–800 kyr [Kent et al., 2017 and references therein] and occurred likely recurrently and at multitude eruption centers along the Atlantic rift basin [Davies et al., 2017]. Potentially, volcanic aerosol enrichments in the atmosphere maintained over a longer period and triggered the cool interval in the upper Langport Member (Figs. 10.2, 10.3, 10.4).

Basaltic magma eruptions, such as the fissure eruptions at Laki in Iceland, usually generate aerosol clouds in the troposphere, reaching the stratosphere only occasionally [Devine et al., 1984; Walker et al., 1984; Palais & Sigurdsson, 1989]. It could be shown, however, that even the Laki eruption, which was several orders of magnitude smaller than those of the CAMP, caused a recognizable cooler period over even several years [Sigurdsson, 1982; Rampino et al., 1988; Thordarson & Self, 2003]. Moreover,
Figure 10.4 Oxygen and carbon isotope values from pristine oysters originating from the earliest Jurassic successions at Lavernock Point, St Audrie’s Bay, and Watchet, United Kingdom, from Korte et al. [2009; including results from van de Schootbrugge et al., 2007] and bulk organic δ¹³C data from Hesselbo et al. [2002], plotted against stratigraphy. Modified after Korte et al. [2009]. Reproduced with permission of the Geological Society of London.
aerosols originating from basaltic eruptions contain 10–50 times more H$_2$SO$_4$ than those from felsic eruptions and influence climate much more effectively [Rampino et al., 1988; Palais & Sigurdsson, 1989]. Taking this latter observations and evaluations into account, the ~2.5‰ decrease in $\delta^{18}O$ (~10 °C warming in the Early Jurassic) would then reflect a shift back from volcanically induced relatively cool temperatures to more normal climatic conditions.

10.4.4. Strontium and Osmium Isotope Stratigraphy

The TJB falls into an interval of a ~30 Myr long gradual decline of the marine $^{87}$Sr/$^{86}$Sr curve which commenced in the Norian (Late Triassic) at a ratio near 0.7080 [Korte et al., 2003]. This decreasing trend was terminated by a rebound to more radiogenic values in the earliest Toarcian (Early Jurassic) after reaching a ratio close to 0.7070 [Jones et al., 1994a]. The finer structure of the marine $^{87}$Sr/$^{86}$Sr curve across the TJB where values are near 0.7077 [Jones et al., 1994a; Korte et al., 2003] is not defined with great confidence at present.

It has been suggested that a rapid decrease of marine $^{87}$Sr/$^{86}$Sr in the Rhaetian was briefly reversed in the latest Rhaetian and in the earliest Jurassic followed by a phase of zero change [Cohen & Coe, 2007] or even to brief increase [Callegaro et al., 2012] that lasted throughout the entire Hettangian stage [Cohen & Coe, 2007]. This interpretation hinges on the accuracy of the geological time scale, the direct comparability of $^{87}$Sr/$^{86}$Sr ratios from Late Triassic Austrian shelf seas to latest Triassic and earliest Jurassic UK shelf seas, and the pristine preservation state of two oyster specimens from the Rhaetian of the United Kingdom. Opposing this scenario stands the current marine Sr isotope curve [McArthur et al., 2012] which adds four Rhaetian conodont values from the Hungarian Csővár section to Rhaetian brachiopod data from the Austrian Weißloferbach section [Korte et al., 2003] but discards earliest Jurassic oyster data from Jones et al. [1994a]. A statistical fit through these data suggests a minor slowdown of the marine $^{87}$Sr/$^{86}$Sr decline during the latest Rhaetian and Hettangian. This slowdown, however, is not constrained by any tie point in the critical time interval of the TJB due to the choice of which fossil materials can be regarded as trustworthy.

Regardless of the adopted model, at some point shortly preceding the TJB, the seawater Sr reservoir appears to have been affected by a more or less substantial shift in the balance of unradiogenic (mantle) and radiogenic (continental) Sr sources to the oceans. This shift has been tied to fluctuations in Os isotope ratios as measured in organic-rich mudrocks in the United Kingdom [Cohen et al., 1999] and related to the emplacement of the CAMP [Cohen & Coe, 2007]. Such a shift in Sr isotope values could either be related to a stronger flux of continental Sr, for example, by way of globally increased weathering rate or by weaker riverine drainage from catchments with unradiogenic $^{87}$Sr/$^{86}$Sr ratios. Alternatively, this effect could also have been brought about by a reduction of mid-ocean ridge activity [cf., Jones et al. 1994b]. CAMP rocks themselves, emplaced at low latitudes and likely highly susceptible to weathering [Cohen & Coe, 2007], would rather have counteracted the observed change in slope of the marine $^{87}$Sr/$^{86}$Sr curve. CAMP weathering would have led to the injection of somewhat more unradiogenic Sr into the coeval seawater, as most reconstructed initial $^{87}$Sr/$^{86}$Sr ratios for CAMP igneous rocks fall within a range from 0.705 to 0.708 [Whalen et al., 2015], slightly lower than the ratio of coeval seawater. CAMP's direct influence on the marine $^{87}$Sr/$^{86}$Sr must thus have been overwhelmed by more potent environmental changes brought about by CAMP emplacement and changes to the Earth surface system that happened simultaneously. Potentially enhanced hydrological cycling as well as elevated sulfuric acid rain may have increased weathering of more radiogenic continental/crustal rocks, delivering relatively Sr to the global ocean with higher $^{87}$Sr/$^{86}$Sr ratios, resulting in an Early Jurassic (Hettangian) plateau in seawater $^{87}$Sr/$^{86}$Sr [Jenkyns et al., 2002 and references therein].

10.4.5. Redox Changes Across the Triassic-Jurassic Transition

Major global change events in Earth’s history are often associated with OAEs, in which significant parts of globally distributed marine basins developed anoxic and/or euxinic conditions [Jenkyns, 2010; Percival et al., 2016]. Much discussion has focused on the causes, consequences, and timing of Mesozoic OAEs and associated changes in global (bio)geochemical cycles, biotic and ecosystem response, and carbon drawdown on local, regional, and global scales [Jenkyns, 2010, and references therein]. The end-Triassic mass extinction and T–J transition interval have generally not been considered to be an OAE, but recent studies do suggest anoxic-euxinic conditions developing at this time. Several studies suggest the increased flux to and/or preservation of organic carbon in marine sediments to explain the deposition of laminated organic-rich black shales described from TJB strata [Wignall, 2001b; Bonis et al., 2010; Clemence et al., 2010; Ruhl et al., 2010; Richoz et al., 2012; Thibodeau et al., 2016]. In addition, organic geochemical analyses of T–J deposits in multiple marginal marine basins suggest the increased abundance of isorenieratane biomarkers, derived from bacteria living in a euxinic photic zone, and gammacerane from the boundary between water masses in a stratified water column [Richoz et al., 2012; Jaraula et al., 2013; Kasprak et al., 2015].
10.5. CONCLUSIONS

The T–J transition is marked by significant changes in δ13C values in organic and inorganic substrates from marine and continental (terrestrial and lacustrine) sedimentary records. A (major) negative CIE predates the TJB by ~100–200 kyr and directly coincides with the end-Triassic mass extinction recorded in marine realms. The observed δ13C signature at that time suggests the release of isotopically light carbon into the ocean-atmosphere system. The temporal relationship between T–J carbon cycle change and the emplacement of the CAMP suggests a potentially causative effect. Different mechanisms of carbon release have been proposed, including (i) carbon degassing directly from CAMP basalts, (ii) thermogenic carbon (methane) release from subsurface organic-rich sediments by intruding sills related to CAMP emplacement, and (iii) methane clathrate release from seafloor sediments in response to initial CAMP carbon release which might be associated with global warming. At this point, contrasts in timing and amplitude observed in T–J δ13C curves between different localities, depositional environments, and proxy records do not allow a full consensus on carbon cycle evolution for the TJB.

Changes in the global exogenic carbon cycle across the T–J transition and related changes in atmospheric and oceanic pCO2 did likely impact local, regional, and global climates, environments, and depositional conditions. Changes in marine δ13O values are suggestive of global warming, and seawater 87Sr/86Sr and 187Os/188Os changes suggest increase in continental weathering rates, while other geochemical and sedimentological markers suggest changes in the redox state of, at least, marginal marine basins. The global warming in the Early Jurassic, however, could also reflect a shift back to normal conditions after a cool interval, the latter triggered by aerosol clouds originating from sulfuric acid exhaled from CAMP volcanism. This sulfuric acid would acidify the rain and could also explain enhanced weathering at that time.

Biostratigraphically and chemostatigraphically well constrained T–J sedimentary records, especially from the open marine realm, are relatively scarce compared to other Early Jurassic, Cretaceous, or Cenozoic global change events (such as OAEs). The T–J, however, stands out as it arguably has one of the best constrained stratigraphic frameworks linking continental LIP emplacement with the marine sedimentary environments.

Recent and increasing interest in the processes controlling events at the T–J transition and the increasing number of marine and terrestrial, and globally distributed, sedimentary records being studied strongly enhance our understanding of the drivers of marine and continental paleoclimatic, paleoenvironmental, and paleobiotic change across this highly enigmatic interval in Earth’s history.

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