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Editorial

Triassic–Jurassic boundary events: Problems, progress, possibilities

1. Problems

As for most geological period boundaries, the Triassic–Jurassic (T–J) transition, ~200 million years ago, was a critical juncture in Earth history during which profound biotic and environmental changes took place. Early comparisons with the end-Cretaceous extinction and the involvement of extraterrestrial impact have now largely, although not entirely, given way to more Earth-bound explanations of events. At the T–J boundary the supercontinent Pangaea, which had dominated the palaeogeographic face of the Earth for the previous ~100 million years, began a fragmentation that has lasted through to the present day. The most obvious manifestation of this process was the production of an estimated two and a half million cubic kilometres of magma with a focus at the centre of Pangaea, and now known as the Central Atlantic Magmatic Province, or CAMP. At more-or-less the same time profound changes took place in the key elements of the biosphere, most notably and obviously in the marine carbonate producing organisms, including those upon which we rely for precise stratigraphic correlation such as ammonites. The case for a dominant volcanic *deus ex machina* now looks incontestable, even if the origin of the volcanism and the precise mechanisms by which environmental changes were driven require much further explanation.

Details of timing are crucial for understanding cause and effect relationships in Earth history, and the lack of a reliable and widely applicable biostratigraphic framework has greatly hampered our understanding of T–J events. It is also plainly the case that in order to reconstruct past events, a physical record of their passing is essential. Here again the Triassic–Jurassic boundary has proved problematic because complete marine sedimentary successions are both few and not very far apart, an observation that has strongly suggested unusually low global sea levels. The relative lack of good marine successions has also delayed the definition of the boundary and the selection of a global stratotype section and point (GSSP); at the time of compilation of this collection of papers decisions had not been made.

In order to facilitate advances in these major issues, IGCP Project 458 was set up in 2001 under the leadership of the editors of this special issue. The project was conceived as multi-disciplinary with the aim of integrating palaeontological, stratigraphical, sedimentological, geochemical, geochronological, palaeomagnetic and mineralogical data from T–J boundary sections globally. Amongst the principal activities we anticipated were: field studies directed towards previously known localities as well as recently or newly discovered ones; compilation of global databases with improved and revised taxonomy, biochronology and palaeobiogeography of major fossil groups, and analysis of patterns of the end-Triassic extinction and Early Jurassic recovery; new radiometric ages and high resolution biostratigraphic correlation to establish a reliable temporal framework; assessment of environmental perturbations and their role in different extinction scenarios using geochemical proxy methods; further studies of the Central Atlantic Magmatic Province and the search for a hypothetical end-Triassic impact to provide clues to the trigger of global environmental change. The overarching view was that reconstruction of the end-Triassic events would use an Earth systems approach to integrate all new findings into the most plausible models.

The papers collected in the present volume individually touch upon many of the areas of study anticipated for IGCP project 458. For convenience we have grouped the papers into four main thematic sections, whilst recognizing that many of them span several of these topics. Some of the most important results in terms of relative timing of events around at boundary are summarized in Fig. 1.

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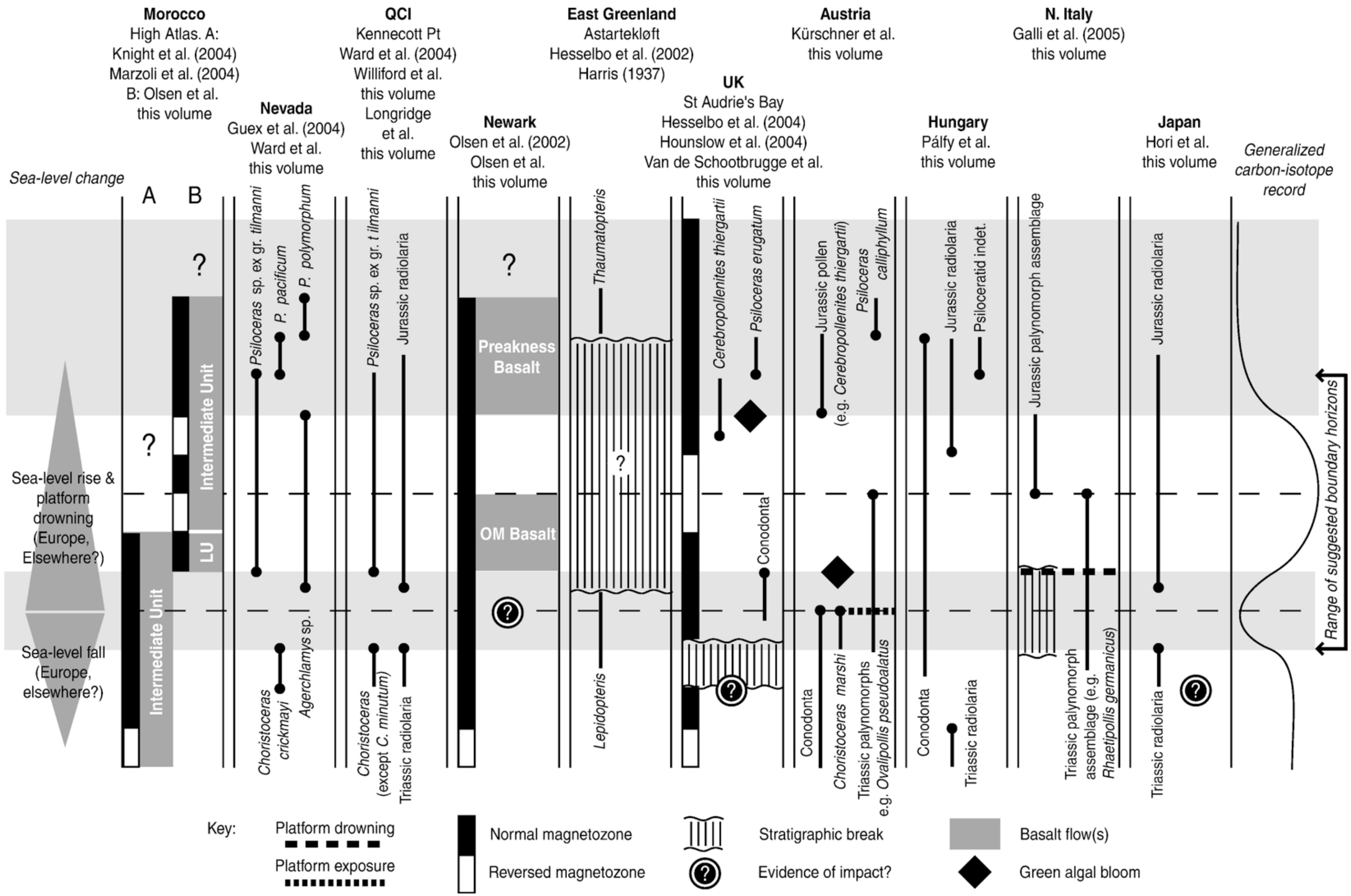

Fig. 1. Relative timing of major events around the T–J boundary, as observed in key sections discussed in this volume. Successions are correlated on the basis of: 1) carbon isotope stratigraphy; 2) ammonite biostratigraphy; 3) radiolarian biostratigraphy, and; 4) magnetostratigraphy. QCI = Queen Charlotte Islands; OM = Orange Mountain; LU = Lower Unit. Comments on *P. tilmanni* from von Hillebrandt (pers comm.).

2. Progress

2.1. The stratigraphic record

The first set of six papers present syntheses of the record of T–J boundary events, most in broadly Tethyan locations. Relatively deep-water settings provide the best opportunities for determination of the sequence of stratigraphic events across the T–J boundary. The carbonate succession at Csővár, Hungary, was deposited in an intra-platform basin that exhibits relatively constant sedimentation through the boundary interval. The section has previously yielded evidence of a negative carbon-isotope anomaly in both bulk carbonate and organic matter co-incident with the palaeontologically defined boundary — as far as this could be identified on the basis of scarce ammonites. In a new study, Pálfy et al. present a truly integrated stratigraphic dataset. The new data lead to important negative and positive findings. Included amongst the most important observations is the rare occurrence of the conodont “*Neohindeodella*” *detrei* 3 m above the first definitively Jurassic psiloceratid ammonite. Although it is difficult to categorically rule out reworking (redeposited beds definitely occur in the succession), the evidence supports the idea that the last conodonts finally went extinct in the earliest Jurassic. Other sedimentary and palaeontological parameters show little change — for example the late Rhaetian clay mineral and foraminiferal assemblages are very similar to those in the early Hettangian. New stable isotope data, carefully screened for diagenetic effects, are also used to suggest that the principal isotope excursion recognized in the succession contains hitherto unrecognized high frequency structure and also preserves a record of significant water mass warming.

Sedimentary archives of three basins from the northern, central and southern Apennines, the La Spezia, the Mt. Camicia, and Lagonegro basins, provide a rich source of information for reconstructing Late Triassic palaeoenvironments and the palaeogeographic evolution of western Tethyan areas. Ciarapica argues that these continuous successions of basinal facies are of particular value, as they also reflect coeval evolution of adjacent platforms through occurrence of platform-derived components. Evidence is inferred for a Late Norian platform drowning, climate change from arid to humid conditions, a spread of dysaerobic facies and increasing eutrophization. The establishment of oligotypic benthic communities, for example suggested by foraminiferan associations dominated by *Triasina hantkeni*, is interpreted as a biotic response and a first step in the end-Triassic extinction. Somewhat at odds with observations from elsewhere, the T–J boundary appears to mark only a second, lesser step of the end-Triassic events. At the T–J boundary the disappearance of the stress-tolerant associations, return of hot and arid climate, and a final, short anoxic episode, followed by rapid resumption of carbonate platform building are observed in the Apennines.

By way of contrast, the Southern Alps of Lombardy, Italy, preserve a T–J transition recorded in a predominantly carbonate shelf and ramp setting. Galli et al. analyze the sedimentary history, faunal and microfloral assemblages, and stable isotope evolution of the boundary interval. The focus of their attention is the newly proposed Malanotte Formation, a conspicuous, thin-bedded, micritic limestone unit that occurs between the fossilifererous, more shallow-water carbonates of the Rhaetian Zu Limestone, and the Hettangian Conchodon Dolomite (cf. Galli et al., 2005). The T–J boundary is drawn near the base of the transgressive Malanotte Formation, on the basis of a gradual change in the pollen assemblages. More pronounced is the slightly earlier abrupt extinction of diverse micro- and macrofaunal associations at the top of the Zu Limestone, a level that is also inferred to represent platform drowning. The T–J boundary is closely correlated with lithological change, accentuated by a gap inferred from an Fe-crusted hardground or a thin layer rich in siliciclastic components. The Malanotte Formation is largely devoid of micro- or macrofauna and lacks the recognizable lithological cycles that characterize the underlying Zu Formation, but it does reveal a three-stage evolution of the carbon isotope ratio in sea water. In the base, i.e. at the T–J boundary, a moderate negative excursion is recorded in bulk organic carbon, followed by a rebound to positive values, which is in turn followed by a more modest negative shift.

Another area with previously less well-known Tethyan T–J boundary sections is the High Tatra Mts in the Western Carpathians, at the Slovak–Polish border. There, the intra-shelf Zliechov Basin, broadly similar to the more familiar Alpine Kössen basins, preserve a record of T–J transition studied in several sections. Michalík et al. present new data from four key sections that were subject to multidisciplinary investigation. Carbonate deposition of the Fatra Formation, consisting of numerous shallowing-upward cycles, was abruptly terminated at the erosional T–J boundary. The conspicuous ‘boundary shale’ of the overlying Kopienec Formation is suggested to reflect both a carbonate production crisis and a sudden increase of riverine influx of terrigenous fine siliciclastic sediment, conceivably related to global changes in climate and ocean chemistry. The moderately diverse Rhaetian biota of the Fatra Formation, largely inferred from skeletal components in the microfacies, disappears at the boundary. A turnover is observed among the foraminifera, which here provide the primary biostratigraphic framework. The earliest Hettangian associations in the Kopienec Formation are dominated by stress tolerant ostracods, which also suggest eutrophic conditions. Immediately below the boundary, a negative carbon isotope anomaly is recorded from bulk carbonate, although it is definitely modest in comparison to some other reported T–J boundary anomalies. Even more ambiguous is the presence of microspheres in limestone beds not far below the boundary. Although they approximately correlate with the level of extinction and isotope anomaly, their origin has not been convincingly demonstrated. However tempting it is to infer impact-related spherules, a more mundane explanation is that they are diagenetic or hydrothermal alteration products of spherical primary sedimentary particles, e.g. ooids.

Depositional environments in the northeastern part of the Iberian Peninsula were quite different from those in the fully Tethyan areas around the T–J boundary. Gómez et al. summarize a wealth of stratigraphic, sedimentological, and palynological information and report new geochemical data. Eustatic sea-level changes variably covered the low-relief area with extensive, extremely shallow carbonate platforms and coastal playa and sabkha flats, leaving a stratigraphic record of mixed carbonates and evaporites arranged in sedimentary cycles. Contrary to earlier opinions, Gómez et al. find no evidence for any major sea-level change or unconformity near or at the T–J boundary, and emphasize the remarkable lateral continuity of the latest Triassic and earliest Jurassic strata. Asturias is the only area of Iberia with a slightly different record: the T–J boundary is placed in a carbonate sequence there, whereas elsewhere it falls within an evaporitic unit. The carbonates in Asturias yield an organic carbon isotope record and add a new entry to the growing list of locations where the T–J boundary negative $\delta^{13}C$ anomaly is recognized. The boundary is drawn with varying precision on the basis of palynology, and the distribution of palynomorphs is also used to infer a climate and vegetation history. A moderate latest Rhaetian plant extinction is followed by considerable diversification in the earliest Hettangian, which appears related to a shift from arid to warmer and more humid climate conditions, as reflected by a change in dominance of xerophytic to hygrophytic pollen producers.

The continental record of the T–J events is no less important than the marine record. Tanner and Lucas provide a significant re-interpretation of the facies and stratigraphic relationships among the upper part of the Chinle Group and the lower part of the Glen Canyon Group of Utah and Arizona, which were situated near the western margin of Pangaea at the time. Their discussion centres on the development of erg deposits within the Wingate Formation, which initiated in the latest Triassic, and which were perhaps fuelled by exposed shoreline sands during a Rhaetian lowstand in sea-level. Their main conclusion is that the mosaic of continental facies is best deciphered in the context of a north–south palaeogeographic transition from dominantly fluvial–lacustrine environments in the north, to dominantly aeolian palaeoenvironments in the south. Reinterpretation of formation boundaries results in the recognition of more than one regional unconformity between demonstrable Triassic strata of the Chinle Group and the Jurassic Glen Canyon Group. This has significant implications regarding a precise placement of a T–J boundary in these often poorly fossiliferous rocks.

2.2. Biotic change

There are many examples of major environmental change events in the Phanerozoic that are characterized by the flood abundances of opportunistic, or ‘disaster’ taxa, but their presence has not hitherto been highlighted for the T–J boundary event. Van de Schootbrugge et al. examine the stratigraphic micropalaeontology of the candidate GSSP section at St Audrie’s Bay, England, and quantify changes in both organic walled and calcareous microfossils at the start of the ‘main’ negative isotope excursion (i.e. the long duration shift to light carbon isotope values that occurs at St Audrie’s Bay about 2 m below the lowest examples of the Jurassic ammonite *Psiloceras*). In the St Audrie’s Bay section it is shown that members of the green algae – prasinophytes and acritarchs – become particularly abundant at the onset of the ‘main’ negative excursion; at the same time, red algae and calcareous nannoplankton are minor constituents of the microflora. The observations are interpreted by Van de Schootbrugge et al. to represent an ecosystem response to raised atmospheric CO_2. Isotope and elemental data from the oyster *Liostrea hisingeri*, collected through the same interval, provide valuable indications of parallel changes in major environmental variables. These data incidentally provide the first convincing evidence that the ‘main’ T–J carbon isotopic curve based on bulk organic matter is present in marine carbonate as well, albeit with half the amplitude (in common with other Mesozoic excursions). Oxygen isotopes and Mg/Ca ratios from the oysters are used to argue for a 4 °C sea-floor temperature increase, and a parallel decrease in salinity by at least 3 PSU at the start of the ‘main’ negative isotope excursion.

The Queen Charlotte Islands of northwestern Canada continue to provide rich palaeontological data, central to our understanding of the T–J extinction. Longridge et al. add new data on the ammonoid and radiolarian diversity trends and biochronology of two important T–J boundary sections in the Queen Charlotte Islands. Of these sections, the one on Kunga Island, has previously provided the sole radiometric age estimate for the T–J boundary in marine rocks, and the other, at Kennecott Point, has yielded one of the best carbon isotope records spanning the extinction interval. Longridge et al. document a moderately diverse ammonoid succession across the boundary, including new discoveries that significantly reduce the ammonoid ‘gap’ of the boundary interval and now permit correlation to early Hettangian ammonoid zones recognized at New York Canyon, Nevada. Not to be overlooked is the excellent radiolarian record from this interval in which Longridge et al. describe a profound decrease in not only radiolarian diversity, but also morphologic complexity amongst earliest Jurassic spumellarian and entactiniid taxa.

To measure a mass extinction only by the proportion of lost taxa and change in diversity is an oversimplification. The ecologic impact may be equally significant and can be estimated by assessing the reorganization of communities. The compositional change of brachiopod communities across the T–J boundary in the Northern Calcareous Alps is investigated by Tomašových and Siblík. Using an array of multivariate analytical techniques, they demonstrate the profound effects among brachiopods during the T–J boundary extinction. The turnover at the boundary is an order of magnitude higher than within the Rhaetian and the Hettangian. Contrary to some earlier suggestions, the T–J brachiopod extinction is abrupt, with no indication of any protracted decline during the Rhaetian. Removal of the incumbents, i.e. extinction of superfamilies with dominant members in latest Triassic communities, led to a fundamental reorganization of community structure. Testing for two competing hypotheses, Tomašových and Siblík find more support for true compositional change across the T–J boundary than they do for a previous proposal of changing habitat preference within major brachiopod groups. Brachiopods are rare, but not absent, in the earliest Hettangian survival phase. Their recovery was underway by late early Hettangian to mid Hettangian times, as indicated by newly established communities with an increasing degree of between-habitat differentiation.

An alternative way to analyze extinction characteristics is to interrogate a global database. This approach has the advantage of providing an overview, with the disadvantage of reduced stratigraphic resolution. Kiessling et al. use the Paleobiology Database to analyze abundance and diversity patterns of marine benthic organisms (sponges, corals, bivalves, gastropods and brachiopods) from the Middle Triassic (~240 Ma ago) to the Middle Jurassic (~160 Ma ago), paying particular attention to possible biases in the dataset. Their analysis confirms the reality of the T–J mass extinction, but it also throws up some evidence for selectivity for certain groups. Taxa that were reef-dwelling, with an inshore habitat preference, preferring carbonate substrates, and confined to low latitudes, exhibit higher extinction risk than other groups. Intriguingly, the same characteristics seem also to apply to background extinctions, lending weight to the idea that the T–J extinction represents an intensification of background processes with, perhaps, an emphasis on extinctions in reefs and inshore environments during (or at the end of) the Rhaetian.

Where body fossils are absent, trace fossils might provide crucial additional information about extinction patterns. An analysis of the T–J boundary trace fossil record is provided by Barras and Twitchett for three sites in southern England, including the candidate GSSP at St Audrie’s Bay. This contribution provides a detailed account of changing ichnofauna of an interval from the upper Langport Member of the Lilstock Formation through five Jurassic ammonoid zones of the Blue Lias Formation (culminating in the *semicostatum* Zone). Their data reveal how eight ichnogenera show significant patterns of infaunal changes through the interval. Above a moderately diverse ichnofossil assemblage in the Langport Member is a notable gap in trace fossils in the ‘Pre-Planorbis Beds’. The authors do not relate this absence of ichnotaxa directly to CAMP effects, because of perceived differences in timing, but instead point up a role for marine anoxia. The focus of their study is the Early Jurassic recovery interval, rather than the lead up to the extinction. The recovery amongst ichnotaxa above the Pre-Planorbis Beds documents a significant increase in ichnotaxic diversity and an increase in the depth of burrowing.

Complementary to the well-known continental sequences in eastern North America are those of the western United States: the vast outcrops of fluvial, aeolian, and lacustrine sedimentary rocks of the Chinle and Glen Canyon groups. In a companion paper to their stratigraphic account, Lucas and Tanner document what is perhaps the best known terrestrial vertebrate record spanning the T–J boundary, including reptilian skeletal remains as well as their traces. They provide a revised biochronology for the interval and subdivide the Late

Triassic and Early Jurassic strata into five biochrons based upon the first appearance of reptile taxa. A significant finding is an increase in both the abundance and size of dinosaurian ichnotaxa leading up to the T–J boundary. This event corresponds to the loss of cruotarsan and phytosaur reptiles and the footprint ichnogenus *Brachychirotherium*.

2.3. Carbon-isotope stratigraphy

The precise stratigraphic relationship between biostratigraphically important fossil groups and carbon–isotope compositions of carbonate and organic sedimentary matter has become critical to understanding T–J events, as emphasized in Fig. 1. In an integrated palynological and isotopic study of the classic boundary sections of the Salzkammergut, Austria, Kürschner et al. provide answers to several outstanding questions of correlation. By constructing a composite carbon isotope curve of bulk organic matter from two nearby sections, they find the now increasingly replicated pattern of an abrupt 'initial' negative isotope excursion, closely followed by an extended 'main' isotope excursion (Hesselbo et al., 2002, 2004). The initial isotope excursion occurs immediately above the top of the hemipelagic carbonate Kössen Formation, in the lowest few centimetres of the 'Grenzmergel' (or boundary marl), and it had been missed in a previous isotopic study of an adjacent section due to relatively wide sample spacing. The negative excursion is coincident with the highest occurrence of conodonts, and the succeeding 1–2 m sees the highest occurrences of typically Triassic palynomorphs. The start of the 'main' isotope excursion occurs at the same level as the lowest occurrence of *Cerebropollenites thiergartii*, a pollen grain that has previously been suggested as a base-Jurassic marker. Whatever taxon is adopted as a definitive guide for the T–J boundary, it becomes clear that the principal period of environmental change takes place within the Grenzmergel and is bracketed by the two negative isotope excursions. Interestingly, like Van de Schootbrugge et al., Kuerschner et al. also recognize the occurrence of a green-algal bloom, but in this case at the same time as the initial negative excursion.

The candidate GSSP at Muller Canyon, Nevada, USA, is another crucial section that reveals the relationship between the organic carbon–isotope curve and biostratigraphically important taxa — in this case ammonites and bivalves. In a re-sampling and re-measuring exercise, Ward et al. reproduce the broad characteristics of a previously published carbon–isotope curve based on bulk marine organic matter (Guex et al., 2004). However, they also find important contrasts with the previous work. Most notably, Ward et al. recognize that the lowest occurrence of the typically Jurassic pectinacean bivalve *Agerchlamys boellingi*, and the lowest find of the ammonite *Psiloceras* sp., occur immediately above an 'initial' negative isotope excursion as defined by multiple data points. If the carbon–isotope curve can be relied upon for correlation, which looks increasingly likely, then the implication is that base of the Jurassic as defined in North American sections on any faunal criterion correlates to horizons many believe to be Triassic in European sections.

In addition to yielding an important record of biotic change across the T–J interval boundary the Queen Charlotte Islands' succession in Canada was one of the first to show evidence for an abrupt negative carbon–isotope excursion coincident with biotic change, in this case radiolarians. Williford et al. here present an extended record of carbon isotope data from bulk organic matter from the Hettangian succession at Kennecott Point in the Queen Charlotte Islands (cf. Ward et al., 2004). A really striking feature of their new data is the magnitude of a positive excursion lying between an initial negative excursion (corresponding closely to the level of radiolarian turnover) and what they interpret as the main (Hettangian) negative excursion. Explanations of the T–J boundary record now have to include both a potential source of isotopically light carbon to produce the negative excursion and an explanation for where all the light carbon subsequently goes. Williford et al. prefer a scenario that involves principally a switch of carbon burial flux from carbonate to organic matter.

2.4. Causes and consequences

Plate motions incessantly operate in the background of all other Earth phenomena. The changing palaeogeography around the T–J boundary is analyzed by Golonka, on the basis of two global palaeogeographic maps constructed for the Late Triassic and Early Jurassic, respectively. More detailed lithofacies maps for the two intervals are provided for crucial areas where the T–J transition proved eventful, including the western Tethys, eastern Tethys, Palaeotethys and eastern Asia, north-western Laurasia, and western Gondwana. The closure of Palaeotethys was expressed in the main convergent event, the Indosinian orogeny, which completed the assembly of eastern Pangaea. In the same time, rifting in the future Central Atlantic area heralded the break-up of the supercontinent. The changing palaeogeography is an important backdrop to the T–J boundary events but most tectonic phenomena operate at longer time scales. A notable exception is the magmatism of the Central Atlantic Magmatic Province (CAMP).

Indeed, flood basalt volcanism of the CAMP is implicated in the currently most favoured scenario explaining environmental changes and biotic extinctions at the T–J boundary. Clearly, relative timing of the boundary events and the eruptions, and the duration of the latter, is of paramount importance in refining or refuting the purported causal link. Two sister papers in this volume contribute new radio-isotopic ages for CAMP basalts and interpret their significance.

Vérati et al. present a suite of 12 new $^{40}Ar/^{39}Ar$ ages from Moroccan CAMP basalts complemented by another two ages from correlative lava flows from Portugal. In Morocco, the CAMP flows are grouped into four units on the basis of their stratigraphy and geochemical characteristics. The first three flow units account for 90% of the total lava volume. Significantly, their ages overlap within error, suggesting that the bulk of volcanic activity occurred within a short time span, in less (perhaps much less) than the 2 Ma resolution afforded by the analytical uncertainty of the dating method. The Moroccan ages are centered around a mean of 199.1 ± 1 Ma ago. Only the fourth and volumetrically minor flow unit has a resolvably youngest mean age of 196.6 Ma ago. The flow ages and chemical compositions suggest that this unit is a product of late-stage asthenospheric upwelling, representing a milestone in the magmatic evolution of the Atlantic rifting process. The Portuguese lava flows are demonstrably coeval with their Moroccan counterparts and unquestionably can be assigned to the CAMP. Significantly, the new suite of ages presented here confirm the earlier suggestion that CAMP volcanism is synchronous with the T–J boundary. The caveat is a recognition of problems associated with both the $^{40}Ar/^{39}Ar$ method applied here and the U–Pb method used to date the boundary from an ash bed in a marine section.

Nomade et al. set out to address the same problems: what is the chronology (i.e. age and duration) of CAMP volcanism and, on the basis of the temporal relationships, how is it related to the T–J boundary events? The team also reports a set of new $^{40}Ar/^{39}Ar$ ages from their 17 samples, split among three of the four continents where CAMP occurs. This brings the total number of published dates to over 100, making CAMP the temporally best constrained large igneous province. Despite chronologic reviews published as recently as in 2003 and 2004, a new effort is justified as some 50 new $^{40}Ar/^{39}Ar$ dates were obtained in the last 3 years alone. The 'quality control' applied by Nomade et al. is also more stringent than in previous studies. After filtering out less reliable dates and those exhibiting disturbed isotopic systems, only the most robust plateau ages are considered further and 58 dates are accepted as valid. It is reassuring that this much larger dataset principally confirms and refines the conclusions of earlier studies. The new synoptic chronology of CAMP reveals that intrusive magmatism commenced ~201 Ma ago, extrusions occurring about 1 Ma later in the African margin, and followed soon after in North America, before spreading to South America. Peak activity, represented by ~80% of the dates, is restricted to a short period between 199 and 197.5 Ma ago. Small-volume eruptions form a protracted tail-end of activity to as late as ~190 Ma ago. A pattern of north-to-south migration of volcanism emerges, although geographic distribution of the data is uneven with the strongest representation of African (mostly Moroccan) samples.

The difference in timing of CAMP volcanism in North America and in North Africa is a matter of some considerable debate (e.g. Knight et al., 2004; Marzoli et al., 2004). Whiteside et al. frame the questions in terms of synchronism between Moroccan and North America activity, and the age relationship to the major pulse of extinction in continental settings, and they attempt to answer these questions using a variety of stratigraphic arguments. Additionally, they provide new cyclostratigraphic, lithostratigraphic, and biostratigraphic data from several continental basins in eastern North America and Morocco. Significant are the new data from Partridge Island (Fundy Basin, Nova Scotia) and the Argana Basin (Morocco), and revised sections elsewhere in North America (e.g. Newark and Hartford basins). On both continents, the authors define an end-Triassic extinction event based primarily on palynology and, to a lesser extent, on tetrapod footprint data. The loss of pollen species and tetrapod ichnotaxa coincides more or less with the onset of *Corollina* (i.e. *Classopollis*) dominated pollen assemblages.

As previously reported, based on astrochronology, the extinction event is proposed to predate the earliest CAMP flow by ~20 ka (e.g. Olsen et al., 2002). Existing basalt geochemical data are used to support this correlation, and Whiteside et al. note that the stratigraphically lowest flows from North America are geochemically High Titanium Quartz normative (HTQ) basalts that are most similar to the HTQ-type flows from the Argana Basalt in Morocco. However, correlation of the North American basalts to the Central High Atlas Basin in Morocco is problematic as these are High Iron High Titanium Quartz normative (HFTQ) basalts for which there are no real correlatives in North America. Whiteside et al. propose that the HFTQ flows of the central High Atlas Basin are part of a magmatic sequence in which the HFTQ evolved from earlier HTQ magmas. Additionally, they specifically dispute a previous correlation of the short reverse magnetochron recognized in Morocco which had implied that North American flood basalts are younger than those found in Morocco. Instead, they suggest that a short reverse magnetochron may yet be found in poorly sampled North America basalts above the palynologically defined T–J boundary, and propose that an independent test of their hypothesis would be recognition of the 'initial' carbon–isotope negative excursion in strata below the oldest basalts in these continental settings.

Ocean acidification, through the build up of dissolved carbon dioxide in the oceans, has been an important putative mechanism behind degradation of marine carbonate ecosystems for several past events (as well as at the present day). This is particularly relevant for times when carbonate platform drowning appears to have accelerated, when extinctions take place preferentially within shallow marine carbonate communities, and when carbonate skeletal mineralogy seems to undergo significant change. Berner and Beerling apply a numerical carbon cycle model to investigate whether volcanic gases of direct magmatic origin were sufficient in quantity to account for these phenomena via oceanic carbonate undersaturation. In addition to the role of carbon dioxide, they also examine the part played by sulphur dioxide, and the possible relative amounts of these two gases during basaltic volcanism, together with feedback mechanisms that potentially include release of methane from gas hydrates. Their conclusions are simple; gasses directly produced from CAMP volcanism can explain oceanic carbonate undersaturation phenomena, but only just. It is necessary to have starting conditions close to undersaturation (i.e. very high atmospheric carbon dioxide) and release of amounts volcanic gas at the very upper limits of plausibility.

CAMP is implicated not only in the generation of excess atmospheric and oceanic carbon dioxide, but also in its drawdown via carbonation reactions during weathering. The seawater record of 'signature' isotopes such as strontium and osmium, which are biased towards unradiogenic values in juvenile basalts, may give a clue as to how CAMP affected weathering processes. Cohen and Coe compile parallel Sr and Os isotope datasets from across the T–J boundary and carry out a semi-quantitative analysis of the results. They find that close similarities exist between the Sr and Os isotope records of the T–J boundary and those of the Toarcian Oceanic Anoxic Event, some 17 million years later, which also coincided with eruption of a continental flood basalt Large Igneous Province (LIP), the Karoo–Ferrar. Perturbations to the seawater Sr-isotope record coincident with LIP emplacement take the form of sudden increases in the proportion of radiogenic strontium, interpreted as increases in continental weathering rates superimposed on an overall trend brought about by long-term decreasing in $^{87}Sr/^{86}Sr$ ratios, presumably reflecting long-term decreasing continental weathering rates. The seawater osmium isotope records for both events also show abrupt changes to more radiogenic values, albeit much more transient than for strontium. Thus, it appears from the T–J boundary record that CAMP eruptions initially promoted a large increase in continental weathering, without the lavas themselves being strongly weathered and contributing a significant unradiogenic flux (the same is also true for Karoo–Ferrar). In the case of CAMP, the subsequent Os-isotope record suggests that this situation was short lived: a rapid return to unradiogenic Os values in the earliest Hettangian indicates input of Os directly from the intense weathering of CAMP lavas that lasted for the next ~3 Ma. By the end of the Hettangian it was all over, with both Sr and Os isotope values returning to their long-term trajectories.

Some of the best clues to the end-Triassic events may have been buried deep in Panthalassa. Hori et al. made an attempt to read the palaeontological and geochemical archives preserved in a slowly accumulated deep-sea chert sequence in Japan. The radiolarian extinction is one of the most promising palaeontological markers of the T–J boundary. In the Kurusu section of the Inuyama area, the rapid radiolarian turnover is subdivided into three events (E1 to E3). First go some of the taxa of a diverse Triassic assemblage (E1). No more than 0.5 Ma later there is a wholesale extinction of the remaining species and the origination of a few new Jurassic forms. This E2 event is recorded in a single bed that is estimated to have accumulated in less than 10 ka and is taken as the T–J boundary. Significantly, the E2 event also corresponds to the last occurrence of conodonts (*Misikella posthersteini*). Then E3 is a post-extinction interval characterized by a low diversity fauna of small, spherical spumellarians. The level of E1 coincides with tantalizing geochemical signals. Among the Rare Earth Elements, a distinct Ce anomaly is interpreted to signal a brief acidification of sea water. The next higher chert bed records an anomalously high abundance of Platinum Group Elements (PGEs). The Ce anomaly is compatible with CO_2 and SO_2 emissions from either a volcanic or an impact source. However, the PGE peak can be best accounted for by a calculated 2.5% admixing of impact melt-derived material. If this is correct, the putative impact may have played a role in the plankton extinction at E1, but it cannot be directly implicated in the T–J boundary extinction, half a million years after. Significantly, in between lies another chert layer that contains basaltic glass and lithic fragments. If derived from a CAMP source, this may be the first direct evidence that some CAMP eruptions were violent enough to spread airborne volcanic particles around the globe. The Kurusu section clearly yields important pieces of the T–J puzzle, yet fitting them together is not straightforward.

Similarly puzzling is a uniquely extensive metre-scale horizon of soft sediment deformation which occurs immediately below the initial carbon isotope excursion in eight discrete sedimentary basins in the UK region, and covering an area of >250,000 km^2. Simms reviews published evidence for this 'seismite' and concludes that it represents only a single shock event, and is at least locally overlain by sedimentary facies of plausible tsunami origin. The facies successions are closely comparable to those described from shallow marine strata in proximity to the end-Cretaceous Chicxulub impact crater. In view of the great distance from the T–J boundary 'seismite' to the nearest CAMP volcanic rocks, Simms rejects the idea that these beds originated in relation to violent CAMP eruptions. Instead he suggests that the observed phenomena are compatible with an impactor of relatively modest dimensions, possibly some 2–3 km across, forming a so far undiscovered crater of 40–50 km diameter, too small to have had a significant effect on biotic change.

3. Possibilities

Despite much progress, a sufficiently high-resolution geochronological framework is still lacking to firmly establish the temporal link of CAMP's first and/or largest eruptions, the environmental events, and the extinction. The main unresolved issue is the comparison of $^{40}Ar/^{39}Ar$ and U–Pb dates. The first method is used extensively in dating CAMP basalts but suffers from uncertainty in the decay constant of ^{40}K. A current revision of the constant (Villa and Renne, 2005) may require recalculation of all $^{40}Ar/^{39}Ar$ ages and their upward adjustment by ~1% (i.e. a published age of 200 Ma would be in fact be close to 202 Ma). Curiously, the U–Pb method may also have produced ages that systematically err on the young side. The going estimate of the T–J boundary age hinges on a multi-grain zircon U–Pb age (199.6±0.4 Ma, Pálfy et al., 2000). Multi-grain analyses are prone to leave slight Pb loss undetected, hence producing marginally younger ages. The remedy is now available, analysis of individual crystals of zircon, also using improved methods to eliminate the effects of Pb loss (e.g. Mundil et al., 2004). Significantly, a single-crystal $^{206}Pb/^{238}U$ age of 201.27±0.27 Ma has been obtained for the North Mountain basalt, a CAMP flow in Nova Scotia, Canada (Schoene et al., 2006). Only the application of these recent advances will help compare the timing of CAMP and T–J boundary events with greater confidence.

There has been some attempt to use cyclostratigraphy to calibrate the duration of events at the T–J boundary, notably with respect to the CAMP volcanism in eastern North America, but so far cyclostratigraphy has not been used effectively to help understand the marine sections. This is partly because most of the marine sections investigated so far show major facies changes across the boundary, and yet early attempts to use this method have not been entirely unsuccessful (cf. Weedon et al., 1999). With the development of high resolution lithological and chemostratigraphic datasets much future progress should be possible.

The proxy record for atmospheric carbon dioxide change (e.g. McElwain et al., 1999; Tanner et al., 2001) remains relatively weak, and there is much scope for further work in this area, based on analyses of the well-preserved plant fossils and soil carbonates that abound in several basins around the world (e.g. Harris, 1937). An improved terrestrial–marine correlation is essential. One potentially powerful approach that has not yet been harnessed for the T–J boundary, is the use of compound specific carbon–isotopes as an alternative to analysis of bulk organic matter, to provide a carbon–isotope stratigraphy where the effects of mixing of different organic components can be better controlled.

Whatever the quality of the present proxy record, there does now seem to be widespread agreement that carbon dioxide produced directly from CAMP, even with a gas hydrate supplement brought about by greenhouse warming, may not have been enough to cause all of the evident environmental impacts. However, it has been pointed out for other LIPs that baking of organic rich rocks may generate massive additional amounts of atmospheric and oceanic carbon (Svensen et al., 2004; McElwain et al., 2005) and this mechanism remains an unexplored possibility in the case of CAMP. Certainly the huge extensional basins into which CAMP magmas were intruded were at times enriched in organic matter and might have provided a ready substrate for production of thermogenic methane.

The debate about extraterrestrial versus volcanic drivers for environmental change has not yet been concluded, and it is noteworthy that all of the candidate indicators of extraterrestrial impact – reports of PGE's and soft sediment deformation – occur shortly prior to

CAMP volcanic activity. Pure coincidence aside, this observation keeps alive the idea that there is an 'impact signal'–LIP connection, even if the mechanisms remain highly controversial; for example, impact decompression melting, as recently articulated by Elkins-Tanton and Hager (2005), or lithospheric gas explosion (Phipps Morgan et al., 2005).

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