Correlating the end-Triassic mass extinction and flood basalt volcanism at the 100 ka level

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ABSTRACT

New high-precision U/Pb geochronology from volcanic ashes shows that the Triassic-Jurassic boundary and end-Triassic biological crisis from two independent marine stratigraphic sections correlate with the onset of terrestrial flood volcanism in the Central Atlantic Magmatic Province to <150 ka. This narrows the correlation between volcanism and mass extinction by an order of magnitude for any such catastrophe in Earth history. We also show that a concomitant drop and rise in sea level and negative $\delta^{13}C$ spike in the very latest Triassic occurred locally in <290 ka. Such rapid sea-level fluctuations on a global scale require that global cooling and glaciation were closely associated with the end-Triassic extinction and potentially driven by Central Atlantic Magmatic Province volcanism.

INTRODUCTION

Mass extinctions reflect important interactions between biology, geology, geochemical cycles, and climate. The end-Triassic mass extinction is one of the five largest extinctions in Earth history, though considerable uncertainty remains in terms of its duration, causes, and effects. Many workers suggest that the extinction was related directly or indirectly to adverse climate following the onset of the Central Atlantic Magmatic Province (CAMP), which erupted $>2.5 \times 10^6$ km^3 of basalt, possibly in <1 Ma, making it perhaps the most voluminous flood basalt sequence of the Phanerozoic (Marzoli et al., 2004; McHone, 2003; Nomade et al., 2007; Whiteside et al., 2007). However, there remains a need for precise and accurate geochronology to correlate the onset of CAMP volcanism, recorded uniquely in terrestrial sections, with the well-documented marine extinction event (Marzoli et al., 2008; Tanner et al., 2004; Whiteside et al., 2007). Also lacking are time constraints for the rates of the Triassic-Jurassic boundary extinction and associated geochemical and paleoenvironmental fluctuations. We sampled three volcanic ash beds bracketing the Triassic-Jurassic boundary from the Pucara basin, northern Peru (Fig. 1A; Schaltegger et al., 2008), and also the first discovered ash bed from the New York Canyon, Nevada, which has been proposed as the Global Boundary and Stratotype Section and Point for the Triassic-Jurassic boundary (Guex et al., 2004). We also provide new U/Pb zircon data from two labs for the North Mountain Basalt, the lowest CAMP basalt from the Fundy Basin, Nova Scotia (Greenough and Dostal, 1992). Both the ash beds and the North Mountain Basalt were dated using chemical abrasion–isotope dilution–thermal ionization mass spectrometry (CA-ID-TIMS; Mattinson, 2005) U-Pb zircon geochronology employing a new well-calibrated ^{202}Pb-^{205}Pb-^{233}U-^{235}U tracer solution, which removes random uncertainty in mass fractionation during mass spectrometry. Data with this solution are as much as 70% more precise compared to single-Pb/single-U tracers, revealing complexity in tuff zircon populations that require new data interpretation strategies.

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TRIASSIC-JURASSIC BOUNDARY

Recent consensus places the Triassic-Jurassic boundary at the first occurrence of the oldest Jurassic ammonite *Psiloceras spelae*, which marks the beginning of postextinction biodiversity recovery (Guex et al., 2004; Morton and Hesselbo, 2008). Pinpointing the extinction interval is more complicated, but coincides with a sharp negative spike in $\delta^{13}C$ at the end-Triassic, when there were steep declines in the biodiversity of ammonites, bivalves, radiolarians, corals, and conodonts (Morton and Hesselbo, 2008). This initial negative excursion is followed by a gradual positive recovery (Fig. 1B), which precedes a slow negative excursion in the Early Jurassic (beginning at bend N13 in Fig. 1B; Guex et al., 2004; Hesselbo et al., 2004; Kuerschner et al., 2007; Ward et al., 2001). The end-Triassic negative $\delta^{13}C$ excursion is recorded in marine organic and carbonate carbon and continent-derived wood material, illustrating that the anomaly resulted from a global carbon cycle perturbation (Galli et al., 2005; Hesselbo et al., 2004; Pálfy et al., 2001).

Proxies for rising atmospheric CO_2 have been reported from terrestrial fossil plants straddling the Triassic-Jurassic boundary (McElwain et al., 1999; Retallack, 2001), though the effects of other gases such as SO_2 on such proxies may also be important (Guex et al., 2004; Tanner et al., 2007). Terrestrial correlatives to the marine extinction are debated (Lucas and Tanner, 2007; Tanner et al., 2004). An apparent palynological event <1 m below the lowest CAMP basalt in the Newark and Fundy Basins in North America was proposed as correlative of the Triassic-Jurassic boundary (Whiteside et al., 2007); this has been challenged on the basis of biostratigraphic and magnetostratigraphic work from North America and Morocco (Marzoli et al., 2004, 2008). Others argue that vertebrate and palynological biostratigraphy in the Newark and Fundy Basins, respectively, place the Triassic-Jurassic boundary in sedimentary slivers above the North Mountain Basalt (Lucas and Tanner, 2007; Cirilli et al., 2009).

An age for the marine Triassic-Jurassic boundary comes from the Pucara basin in northern Peru, where Schaltegger et al. (2008) reported a weighted-mean ^{206}Pb/^{238}U date of 201.58 ± 0.18/0.38 Ma (2σ; without/with decay constant uncertainties). Abundant CAMP ^{40}Ar/^{39}Ar data cluster at 199 Ma (e.g., Nomade et al., 2007), but uncertainties of 1–2 Ma on individual dates in addition to the well-documented ~0.7%–1% bias between the ^{40}Ar/^{39}Ar and U-Pb dating methods (Kuiper et al., 2008; Schoene et al., 2006) make this correlation imprecise. A ^{206}Pb/^{238}U date from the North Mountain Basalt of 201.27 ± 0.06/0.30 Ma (Schoene et al., 2006) would suggest that the CAMP postdates the Triassic-Jurassic boundary, precluding a causative relationship. However, those two U-Pb dates were measured using different tracer solutions, allowing for systematic bias and preventing high-precision comparison.

LOCALITIES AND U-Pb GEOCHRONOLOGY

We sampled three volcanic ash beds bracketing the Triassic-Jurassic boundary in the Pucara basin (samples LM4–86, LM4–90, and LM4–100/101; Figs. 1A, 1B), which is well calibrated biostratigraphically (Schaltegger et

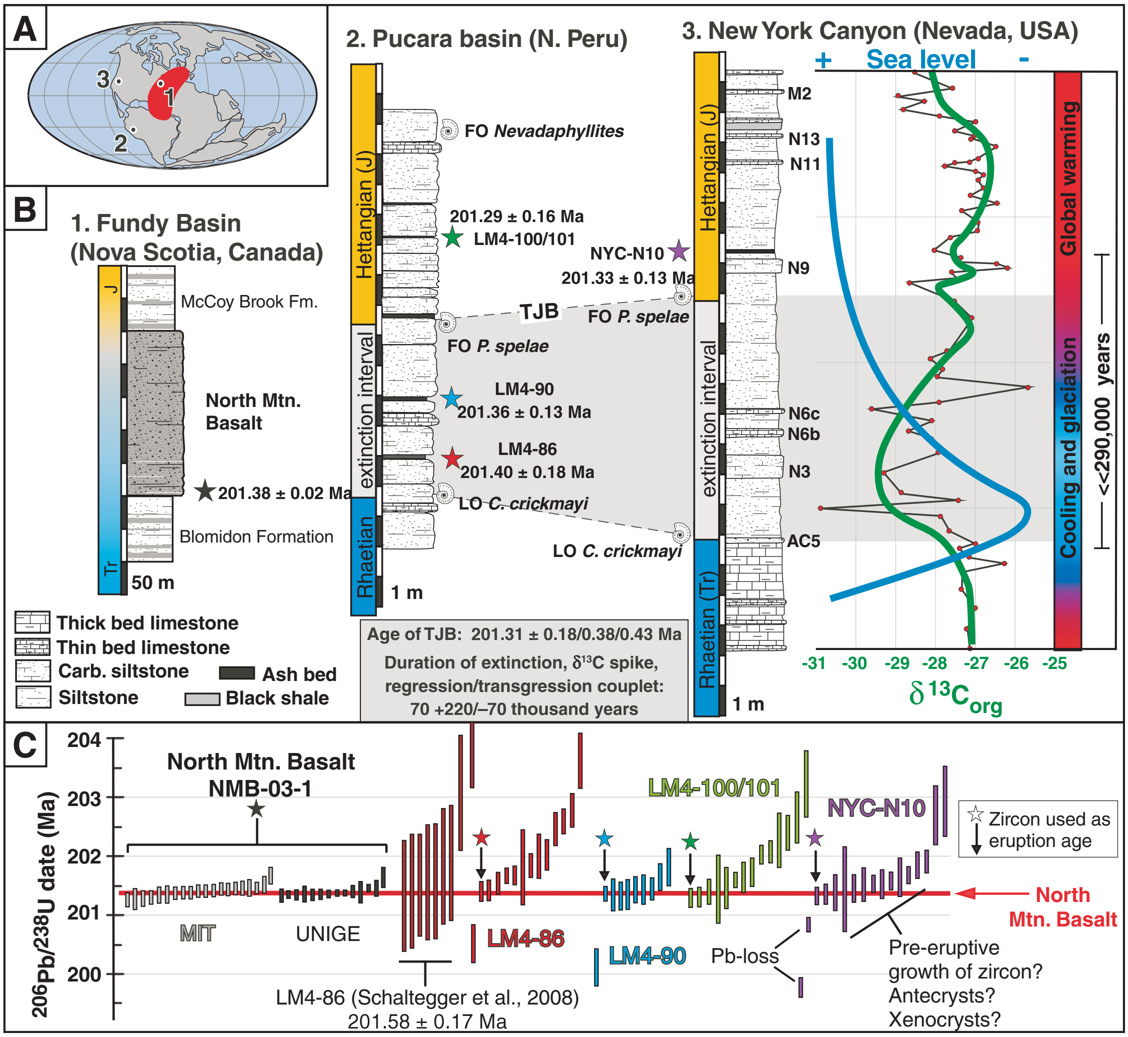

Figure 1. A: Location map of three sections studied, with ca. 200 Ma paleogeography. Numbers correspond to stratigraphic sections in B. Red area outlines approximate extent of Central Atlantic Magmatic Province (e.g., McHone, 2003) B: Stratigraphic columns for sections studied; scale bars at bottom. J—Jurassic; Tr—Triassic. Fundy Basin section is after Whiteside et al. (2007). Pucara basin biostratigraphy is detailed in Schaltegger et al. (2008). New York Canyon stratigraphy, biostratigraphy, bed numbers, carbon isotopes, and sea-level curve are from Guex et al. (2004, 2008). Green curve in $\delta^{13}C$ plot is running mean of red data points. FO—first occurrence. LO—last occurrence; org—organic. Stars indicate ash beds sampled, with interpreted $^{206}Pb/^{238}U$ deposition age. All uncertainties are 2σ. TJB—Triassic-Jurassic boundary, defined as first occurrence of *Psiloceras spelae*. C: $^{206}Pb/^{238}U$ dates for single-grain zircons, color-coded to sample locations in B. Date for LM4–86 from Schaltegger et al. (2008) includes tracer calibration uncertainty. Data for North Mountain Basalt are from Massachusetts Institute of Technology (MIT) and University of Geneva (UNIGE); all ash-bed data are from University of Geneva.

al., 2008). In this section, the disappearance of the latest Triassic ammonite *Choristoceras crickmayi* immediately precedes the peak extinction rate (Guex et al., 2004). One ash bed sample (NYC-N10) was collected ~1.5 m above the first occurrence of *P. spelae* in the New York Canyon section, for which detailed $\delta^{13}C$ and biostratigraphic data have been published (Guex et al., 2004, 2008) (Figs. 1A, 1B). We also provide new U/Pb data from two laboratories for the North Mountain Basalt (NMB-03–1). All dates were produced using CA-ID-TIMS on single zircons relative to the new EARTHTIME (http://www.earth-time.org/) ($\pm^{202}Pb$)-^{205}Pb-^{233}U-^{235}U tracer solution, allowing us to ignore tracer calibration uncertainties within this study (Schoene et al., 2006). All uncertainties are reported at the 2σ level, have been corrected for ^{230}Th disequilibrium, and omit decay constant uncertainties unless otherwise noted (see the GSA Data Repository[1] for analytical details). $^{206}Pb/^{238}U$ dates for single zircons are plotted in Figure 1C and concordia plots and U-Pb data are presented in Figure DR1 and Table DR1, respectively.

Ash beds yielded 20–100 zircons between 50 and 200 μm in diameter (see the Data Repository); 14 zircons from sample LM4–86 give $^{206}Pb/^{238}U$ dates that span ~1.5 Ma between 201 and 203 Ma, with one older and one younger grain. The main population is not statistically equivalent, with a high mean square of weighted deviates (MSWD) of 12 on a weighted-mean $^{206}Pb/^{238}U$ date (i.e., many dates do not overlap at 2σ). Similarly, LM4–90 and LM4–100/101 yield zircon populations that spread between 0.4 and >1 Ma with high MSWD values. LM4–90 also has one grain ca. 200.1 Ma and a population of ca. 900 Ma xenocrystic zircons. Of 17 zircons from NYC-N10, 15 give $^{206}Pb/^{238}U$ dates between 201.3 and 201.9 Ma; a weighted-mean yields an MSWD of >2. Two younger grains give $^{206}Pb/^{238}U$ dates younger than 201 Ma and two older grains of ca. 203 Ma (Fig. 1C).

Analyses of 13 single grains of the North Mountain Basalt (NMB-03–1) from the University of Geneva and 19 analyses from the Massachusetts Institute of Technology are statistically equivalent, yielding a weighted-mean $^{206}Pb/^{238}U$ date of 201.38 ±0.02/0.22/0.31 Ma (internal uncertainties/with tracer calibration uncertainties/with decay constant uncertainties; omitting two analyses, MSWD = 1.2).

INTERPRETING DEPOSITION AGES FOR ASH BEDS

U/Pb ages for volcanic ash beds are often determined by calculating a weighted-mean date and thus assuming that there exists a single population of zircons that crystallized immediately prior to eruption (Ramezani et al., 2007; Schaltegger et al., 2008). Our ash-bed data reveal complicated U-Pb systematics, precluding such

[1]GSA Data Repository item 2010109, U-Pb data table, analytical details, and additional figures, is available online at www.geosociety.org/pubs/ft2010.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.

an interpretation. The observed spread in dates could be created by several effects: (1) analytical bias and underestimated uncertainties, (2) postcrystallization loss of radiogenic Pb from zircons, or (3) zircons representing a range of growth histories prior to eruption. The equivalency of the North Mountain Basalt data from two independent laboratories using a single tracer solution, transparent data-reduction techniques (Schmitz and Schoene, 2007), and careful analytical blank calibration (see the Data Repository) suggests that tuff zircon data are equally accurate and analytical uncertainties are correctly estimated.

Despite the overall success of the chemical abrasion technique at removing the effects of Pb loss (Mattinson, 2005), three of four ash-bed samples have at least one zircon that is far younger than the main population, and we suggest that these outliers are the result of residual Pb loss. There is a simple test suggesting that the main population of zircons did not also undergo Pb loss: the youngest zircon from the main population of each sample is younger or within uncertainty than that of the sample stratigraphically below it. This would be remarkably coincidental if these populations had undergone Pb loss.

We interpret the spread in $^{206}Pb/^{238}U$ dates of the main population of tuff zircons to be the result of protracted growth of zircon in the magmatic system and/or the incorporation of xenocrystic zircon. Complex volcanic and plutonic systems involving small magma batches, mafic and silicic replenishment, magma mingling and/or mixing, and crustal assimilation can entrain antecrystic to xenocrystic zircons with dates recording millions of years of magmatic activity (Charlier et al., 2005; Crowley et al., 2007; Simon and Reid, 2005; Miller et al., 2007). Th/U ratios of zircons from each of our ash samples plot into distinct but overlapping groups (Fig. DR2). This argues that younger ashes came from a different magma batch, but may have incorporated reworked ash material and/or xenocrystic zircons. Zircon morphology supports this: the youngest grains had the highest aspect ratios and were prismatic and euhedral (typical of rhyolitic zircon), whereas older populations contained similarly euhedral grains in addition to anhedral rounded grains. Thus, although zircon selection is critical in avoiding xenocrysts, the increased precision afforded by the ^{202}Pb-^{205}Pb-^{233}U-^{235}U tracer can also identify euhedral zircons that predate eruption. The zircons from LM4–86 dated (Schaltegger et al., 2008) with a ^{205}Pb-^{235}U tracer likely record the same population of zircons as this study, though it appeared as one population due to increased uncertainties on single analyses (Fig. 1C). Such an observation demands caution when taking weighted means of large populations, especially in lower precision data sets.

Therefore, we use the youngest single closed-system zircon to approximate the eruption date (Figs. 1B, 1C). A less conservative interpretation uses weighted means of several zircons, with the resulting MSWD as a guide. Those eruption ages overlap with the single-grain interpretation and with the age of the North Mountain Basalt (Table DR3).

TIMING AND RATES OF EVENTS AT THE TRIASSIC-JURASSIC BOUNDARY

A $^{206}Pb/^{238}U$ age of the Triassic-Jurassic boundary in the Pucara basin, defined as the first appearance of *P. spelae*, can be calculated using our new data to be 201.31 ±0.18/0.38/0.43 Ma, using the difference between the minimum age of LM4-100/101 and the maximum age of LM4-90. This correlates to the onset of the CAMP in the Fundy Basin to within ±150 ka (Fig. 1B).

Our data also provide tie points between the terrestrial Triassic-Jurassic boundary, located near the North Mountain Basalt (Cirilli et al., 2009), and the marine ammonite extinction in two stratigraphic sections. The deposition age from NYC-N10 in Nevada is identical to that from LM4-100/101 in Peru, illustrating that the first appearance of *P. spelae* in both sections is contemporaneous within our resolution. If we assume that the last occurrence of *C. crickmayi* in the New York Canyon is also correlative with that in the Pucara basin, we can conclude that the duration of the end-Triassic negative $\delta^{13}C_{organic}$ excursion was 70 +220/–70 ka (the duration between LM4-86 and NYC-N10), which also serves as an estimate for the duration of the mass extinction event.

DISCUSSION

Some numerical carbon cycle models suggest that CO_2 released from flood basalts is insufficient to create the end-Triassic negative $\delta^{13}C$ anomaly and that associated with the Permian-Triassic extinction (Beerling and Berner, 2002; Payne and Kump, 2007). Alternatively, it is hypothesized that CAMP volcanism may have destabilized methane hydrates or accessed large carbon reservoirs by erupting through organic-rich sediment, resulting in massive input of light carbon into the oceans and atmosphere, creating a <200 ka negative $\delta^{13}C$ spike (Beerling and Berner, 2002; Pálfy et al., 2001; Retallack, 2001), which is consistent with our geochronological data. However, recent work suggests that SO_2 and polycyclic aromatic hydrocarbons were drivers in rapid and widespread terrestrial plant turnover, rather than CO_2 and greenhouse warming (Van de Schootbrugge et al., 2009). Furthermore, Korte et al. (2009) provided $\delta^{18}O$ data from fossil oysters that argue for cool ocean temperatures immediately after the initial negative $\delta^{13}C$ excursion (after the extinction event in the New York Canyon), followed by ~8 °C early Hettangian warming. These recent studies argue that CO_2-induced global warming was not the driver for the end-Triassic biotic crisis, but allow that it was important in the postextinction recovery interval.

The short duration (<290 ka) for the marine regression-transgression sequence in the New York Canyon provided by our geochronological data gives an additional constraint. Latest Triassic regression-transgression is recognized in numerous sections in Europe and North America, and is likely the result of global sea-level change (Hallam and Wignall, 1999). Several sections in Europe also show it coupled to the Late Triassic negative $\delta^{13}C$ excursion, as seen in the New York Canyon (Fig. 1B) (e.g., Galli et al., 2005; Hesselbo et al., 2004; Kuerschner et al., 2007). Global eustasy results from a variety of processes, including continental uplift due to thermal underplating (e.g., by a plume) or changes in volumes or rates of mid-ocean ridge production, but these processes occur on time scales longer than 1 Ma (e.g., Miller et al., 2005). The rapid sea-level fluctuation we document for the latest Triassic can only be explained by glacial eustasy. A model that accounts for this observation was proposed in Guex et al. (2004), and suggested that the negative $\delta^{13}C$ excursion in the uppermost Triassic (Fig. 1B) was associated with extinction and primary productivity collapse caused by volcanic SO_2 and heavy metals emissions, acid rain, and a cooling and glacial event that caused a short but major drop in sea level. This first phase was followed by CAMP–related CO_2 accumulation, greenhouse warming, marine transgression, and postextinction biotic recovery, corresponding to the positive and second negative $\delta^{13}C$ excursions (Fig. 1B).

Fingerprinting the trigger for the end-Triassic ecological disaster must come from additional biological and environmental proxy data in combination with high-precision geochronology in other Triassic-Jurassic boundary sections. Such work will better constrain the rates of CAMP eruption and corroborate that sea-level change and extinction were everywhere fast and contemporaneous (Hallam and Wignall, 1999; Hesselbo et al., 2004). Our new U-Pb data show that such constraints can be facilitated using new freely available EARTHTIME (±^{202}Pb)-^{205}Pb-^{233}U-^{235}U tracers, by effectively eliminating interlaboratory bias and substantially increasing both internal and external precision of U-Pb ID-TIMS dating.

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