The continental Triassic-Jurassic boundary in central Pangea: recent progress and preliminary report of an Ir anomaly

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Abstract

The Triassic-Jurassic boundary marks one of the "big five" mass extinctions in the last half billion years. In many of the exposed rift basins of the Atlantic passive margin of eastern North America and Morocco, the boundary is identified as an interval of stratigraphically abrupt floral and faunal change within cyclical lacustrine sequences. A comparatively thin interval of Jurassic age strata separate the boundary from extensive overlying basalt flows, the "best" dates of which (~202 Ma) are practically indistinguishable from recent dates on tuffs from marine Triassic-Jurassic boundary sequences. The pattern and magnitude of the Triassic-Jurassic boundary at many sections spanning more than 10° of paleolatitude in eastern North America and Morocco is remarkably similar to that at the Cretaceous-Tertiary boundary, sparking much debate on the cause of the end-Triassic extinctions with hypotheses focusing on bolide impacts and climatic changes associated with flood basalt volcanism.

Four prior attempts at finding evidence of impacts at the Triassic-Jurassic boundary in these rift basin localities, were unsuccessful. However, based on much more detailed geochemical and mineralogical sampling of four tightly biostratigraphically-constrained sections in the Newark rift basin (NY, NJ, PA: USA), we have been able to identify a modest Ir anomaly (up to 285 ppt, 0.29 ng/g) measured by gamma-gamma coincidence spectrometry after neutron activation. The Ir anomaly is directly associated with a fern spike, previously identified in these sections. A search for shocked quartz in these samples proved fruitless. Although both the microstratigraphy and the biotic pattern of the boundary is very similar to continental Cretaceous-Tertiary boundary sections in the western US, we cannot completely rule out a volcanic or other non-impact hypotheses for data currently in hand.

Introduction

The Triassic-Jurassic boundary (~202 Ma) marks one of the five largest mass extinctions of the Phanerozoic (Sepkoski, 1997), arguably at least as large in magnitude as that at the much better known Cretaceous-Tertiary boundary (Fig. 1). The extinctions occurred in a hothouse world during a time of extremely high CO_2 (Ekart et al., 1999) and the existence of the supercontinent of Pangea. In continental environments, the Triassic-Jurassic extinctions mark the end of a regime dominated by non-dinosaurian tetrapods and the beginning of the dinosaurian dominance that would last the succeeding 135 million years.

Pangean rift basins developed during the Middle to Late Triassic along a broad zone from Greenland through the Gulf of Mexico in the ~40 million years preceding the Jurassic opening of the central Atlantic Ocean. Many of these rift basins preserve a detailed record of the Triassic-Jurassic boundary in mostly continental environments characterized by relatively high sedimentation rates (Fig. 2). This paper reports on recent progress in documenting the biotic transition around the Triassic-Jurassic boundary, preliminary evidence for an associated Ir anomaly, and the relationship between these boundary events and the Central Atlantic Magmatic Province (CAMP).

Identification of the Continental Triassic-Jurassic Boundary

Rift basins of central Pangea developed largely in a continental milieu. Comparison of the Triassic-Jurassic boundary in the marine realm with that in continental environments is largely based on studies of Alpine marine sections correlated to other European, mostly continental areas by palynomorphs and scant marine invertebrates (Schulz, 1967; Brugman, 1983; Beutler, 1998). Cornet and others (Cornet et al., 1973; Cornet and Traverse, 1975; Cornet, 1977; Cornet and Olsen, 1985) first identified the Triassic-Jurassic boundary in the tropical Pangean basins, most notably in the Jacksonwald syncline of the Newark basin. This palynological work was augmented by subsequent studies in the Newark and other basins in eastern North America by Fowell (Olsen et al., 1990; Fowell, 1994; Fowell, et al., 1994; Fowell and Traverse, 1995).

Within eastern North America, Olsen and Galton (1977, 1984), and Olsen and Sues (1986) identified a transition in the terrestrial tetrapod assemblages that coincides, albeit at a coarser level, with that seen in the palynomorphs. Additional vertebrate paleontological studies have refined the correlation of the Triassic-Jurassic boundary in these continental rifts to the marine realm, in which very rare terrestrial vertebrates are found (Olsen et al, 1987; Huber et al., 1996; Lucas, 1998).

Two significant problems in the biostratigraphic understanding of the Triassic-Jurassic boundary are the interpretation of negative evidence and the presence of distinct biotic provinces during the early Mesozoic. We believe that the Triassic-Jurassic boundary is marked by a catastrophic mass extinction. As in the case of the Cretaceous-Tertiary boundary, we expect the immediate post-boundary biotic assemblages to consist of survivor taxa, not new appearances. However, there is a long tradition of identification of the Triassic-Jurassic boundary based on new appearances, notably the first appearance of the ammonite *Psiloceras planorbis* in marine sections (e.g., Page and Bloos, 1998), and this approach has been carried over to the palynological analyses (e. g. Morbey, 1975).

Arguably, global correlation of a boundary marked by a mass extinction with a very rapid and global cause using the first appearances of taxa is inappropriate. The boundary is more appropriately marked by the last appearances of taxa and not the appearance of new taxa that might appear thousands if not millions of years in the study area after the "event".

A second problem with traditional biostratigraphy across the boundary is a very strong climatic gradient (Kent and Olsen, 2000b). The central European sections that form the basis of much of the palynological and vertebrate biostratigraphy were located some 2000 km north of the tropical Pangean basins of North America on the opposite (northern) side of the northern sub-tropical arid belt (Kent and Muttoni, 2001; Olsen and Kent, 2000). The marked floral and faunal provinciality, probably related to the Pangean climatic belts, was reviewed by Cornet and Olsen (1985) and Olsen and Galton (1984), but the possible effects of this very strong climatic gradient on floral assemblages - particularly palynomorphs - has been either ignored or discounted in attempts at long-distance correlation.

Reliance on first appearances and the discounting of the effects of climatic gradients have led some workers to cite the absence of certain critical palynological taxa characteristic of western Europe and the abruptness of the transition itself as evidence of a major hiatus at the Triassic-Jurassic boundary, especially in eastern North America and Greenland (e.g., Pedersen and Lund, 1980; van Veen et al., 1995; Tourani et al., 2000). It is critical to point out that there is no physical stratigraphic or sedimentologic evidence of such a hiatus in these areas; indeed, detailed cyclostratigraphic and magnetostratigraphic evidence indicates continuous deposition across the palynologically-identified boundary, especially where it has been examined in the most detail in the Newark basin (Kent et al., 1995; Olsen et al., 1996a, 1996b, 2001a, 2001b).

There is no question that there are significant differences between the stratigraphic distribution of palynomorph taxa in tropical Pangean basins and in western Europe. This difference shows up very obviously in the concurrence of the rise of abundance of *Corollina* (*Classopollis*) and persistence of *Patinasporites densus* in eastern North America over the last 15 million years of the Triassic (Cornet, 1977) - a pattern that has no described European counterpart. Indeed, the abrupt disappearance of *Patinasporites densus* is a signature of the Triassic-Jurassic boundary in tropical Pangea. If the same kind of typological biostratigraphic philosophy were applied to the western North American continental

Cretaceous-Tertiary boundary, a major hiatus would be required. Significantly, the Cretaceous-Tertiary boundary also exhibits dramatic climate-related floral provinciality (Sweet et al., 1990).

Radiometric dating of the Triassic-Jurassic boundary provides support for the correct identification of the Triassic-Jurassic boundary in the tropical Pangean rifts. Ages from lavas (and associated intrusive feeders; e.g., Ratcliffe, 1988) just above the palynological Triassic-Jurassic boundary in eastern North America provide U-Pb ages of 201.3±1 Ma (Gettysburg sill, Dunning and Hodych, 1990), 200.9±1 Ma (Palisade sill, Dunning and Hodych, 1990), and 201.7±1.3 Ma (North Mountain Basalt, Hodych and Dunning, 1992) Ma and $^{40}Ar/^{39}Ar$ ages of 202.2±1, 200.3±1.2, and 201.2±1.3 Ma (Palisade sill and Culpeper basin plutons, respectively; Sutter, 1988).

Until recently, these have been the only ages directly applicable to the boundary; however, Palfy et al. (2000a, 2000b) provide a date of 199.6±0.3 Ma for a tuff just below the marine Triassic-Jurassic boundary in British Columbia. Palfy et al. (2000a) argue that the difference between their marine date and the continental dates from eastern North America implies that the continental extinctions occurred prior to the marine extinctions. However, the differences in ages are very small, and thus we regard all these dates as indistinguishable, given reasonable, but unstated, geological and inter-laboratory uncertainties. Further, recent dates on the Orange Mountain Basalt of 201±2.1 Ma (Hames et al., 2000) and its feeder, the Palisade sill, of 201 ± 0.6 Ma (Turrin, 2000), both from the Newark basin, are indistinguishable from Palfy's et al. date (as cited by them). Indeed, the similarity between the marine and continental dates provide powerful support for the age of the Triassic-Jurassic boundary being about ~200 Ma, which is substantially younger than in recent time scales (e.g., Harland et al., 1990; Gradstein et al., 1994).

More precise dating of the boundary requires inter-laboratory cross-calibration and a better understanding of the systematics of the K-Ar system in tholeiites than currently exists (e.g., Turrin, 2000). At the present time we favor a date of 202 Ma for the boundary, based on the assumption that the average of available Newark igneous dates falls close to the middle of the extrusive section, and adding on the duration of the older half of the interval based on Milankovitch cyclostratigraphy, then rounding to the nearest half million years (Olsen et al., 1996b). We believe realistic geological uncertainties for the Newark-based radiometric boundary dates are on the order of 2 million years.

Newark Basin Boundary Sections

Of all of the Triassic-Jurassic Pangean rifts, the Newark basin is arguably the best known stratigraphically, due to over a century of field work, extensive scientific coring by the Newark Basin Coring Project (NBCP) (Olsen et al., 1996a), geotechnical coring by the Army Corps of Engineers (ACE) (Fedosh and Smoot, 1988), and petroleum industry exploration. Virtually the entire >5-km section has been recovered with redundancy in the existing >10 km of continuous core. In addition, the continental Jurassic boundary is better known in the Newark basin than elsewhere. In particular, the boundary is best known in the Jacksonwald syncline in the southwestern part of the basin (Fig. 3), because of ongoing commercial real-estate development and rich fossil content.

It has long been known that much of the Newark basin section is composed of cyclical lake deposits (Van Houten, 1962), and with the recovery of the NBCP and ACE cores, the cyclicity of virtually the entire section has been described (Olsen et al., 1996a, 1996b) (Fig. 4). Van Houten ascribed the hierarchical cyclicity in the Lockatong and Passaic formations of the Newark basin to lake level fluctuations controlled by Milankovitch climate cycles (Van Houten, 1962, 1964, 1969) and his interpretation has proven applicable to the entire cyclical lacustrine sequence (Olsen et al., 1996a; Olsen and Kent, 1996, 1999). Similarly, the magnetostratigraphy of the Newark basin section has been worked out in cores and outcrop (Kent et al., 1995). Taken together, the magnetostratigraphy, Milankovitch cyclostratigraphy, and radiometric dates from the lavas have provided the basis for an astronomically tuned time scale for the Late Triassic (Kent and Olsen, 1999a, 2000a) and earliest Early Jurassic (Olsen et al., 1996b; Kent and Olsen, 1999b) (Fig. 4).

Recent work on outcrops in the Hartford basin (Fig. 1) has resulted in a preliminary magnetostratigraphy and cyclostratigraphy that allows astronomical calibration of the two million years of section post-dating the Jurassic lava flows in the Hartford basins that are precisely correlative with those in the Newark basin (Fig. 4) (Kent and Olsen, 1999b). This interval includes the oldest reversed polarity zones in the Early Jurassic, providing a tie to marine sections in the Paris basin as well as an upper bound to the normal polarity zone enclosing the exposed CAMP lavas.

The Jacksonwald syncline of the Newark basin has an unusually thick latest Triassic-earliest Jurassic section, marked by accumulation rates greater than anywhere else in the basin, and capped by the Orange Mountain Basalt and Feltville Formation (Fig. 3). The cyclicity and paleontological richness of the latest Triassic age sections here are better developed than anywhere else in the Newark basin. A series of largely temporary exposures created for houses over the last 15 years have allowed detailed paleontological and cyclostratigraphic analyses of the uppermost Passaic Formation including the Triassic-Jurassic boundary (Figs. 5-7), and it is these sections that we concentrated on for the new geochemical analyses.

The boundary section exhibits strongly cyclical sediment variation with well-developed gray and black shales occurring periodically (in terms of thickness) in the Milankovitch pattern typical of Newark basin lacustrine sequences. In general, gray strata produce pollen and spores and thus the boundary section is better constrained biostratigraphically here than elsewhere (e.g., northern Newark basin). Over the years boundary sections have been exposed over a distance of about 2 km along strike with the easternmost (i.e., most basinward) exposures being the finest grained and having the highest proportion of gray strata. All of the Newark basin boundary sections are characterized by a litho- and biostratigraphy that is laterally consistent, despite the lateral changes in facies and accumulation rate (Fig. 5).

A very prominent gray shale occurs about 25-30 m below the Orange Mountain Basalt (Figs. 5, 8). This unit contains palynoflorules that are dominated (60%) by *Patinasporites densus* with variable amounts (5-20%) of *Corollina*, and other pollen and spores. This is typical of many older Late Triassic palynoflora assemblages from the Jacksonwald syncline. Above this gray shale are variegated red and gray shales and sandstones that have a lower abundance (5-40%) of *Patinasporites densus*. Approximately 8-12 m below the Orange Mountain Basalt another prominent marker bed occurs: a brown to "blue-gray sandstone" with abundant comminuted charcoal and wood, which we refer to as the "blue-gray sandstone" bed (Figs. 5, 6, 8). This unit generally lies directly above a thin (1-10 cm) coal bed or carbonaceous shale. The highest stratigraphic occurrence of *Patinasporites densus* occurs about one meter below the sandstone (Fig. 9). This palynoflorule (sample 6-2 of Fowell, 1994) is dominated by *Corollina* (73%) with about 10% *Patinasporites*. This is the highest assemblage that we regard as having a Triassic-aspect palynomorph assemblage.

In the 40 cm below the "blue-gray sandstone", palynomorph assemblages are consistently dominated by trilete spores belonging to taxa usually attributed to ferns (Fig. 9) (first discovered by Litwin cited in Smith et al., 1988). Within a couple of centimeters of the base of the "blue-gray sandstone", the proportion of spores reaches a maximum of 80% (Fowell et al., 1994). *Patinasporites densus* is absent from these assemblages, and we consider these assemblages to be of earliest Jurassic age. Such fern-dominated assemblages are unknown elsewhere in Newark basin strata of Triassic age. We refer to this very anomalous high proportion of fern spores as a "fern spike", in parallel with the terminology used for the Cretaceous-Tertiary boundary (e.g., Tschudy et al., 1984; Nichols and Fleming, 1990). It is within this fern spore-rich interval that the iridium anomaly occurs (see below). Above the "blue-gray sandstone", palynoflorules are dominated by *Corollina* whilst *Patinasporites* and other Triassic-type taxa are absent. The character of these assemblages is similar to that of the Jurassic strata overlying the basalts throughout the Newark Supergroup, and we consider them to indicate a Jurassic age for the uppermost Passaic Formation. Thus, again in analogy to the Cretaceous-Tertiary boundary, we hypothesize that the fern spike at least approximates the base of the Jurassic, and hence the Triassic-Jurassic boundary (Fig. 7). Based on the cyclostratigraphically-based accumulation rate, the last Triassic aspect palynoflorule occurs, conservatively, within 25 ky of the base of the Orange Mountain Basalt, the interval of time between this Triassic assemblage and the lowest definitive Jurassic assemblage (Jb-6 of Cornet, 1977) is less than 10 ky. Similarly, the Triassic-Jurassic boundary, as defined here by the fern spike, occurs within 20 ky of the base of the Orange Mountain Basalt.

Reptile footprint taxa broadly follow the same pattern as the palynological stratigraphy. Footprint faunules have been recovered at many levels within the Jacksonwald syncline and there is a concentration of productive levels near the palynologically-identified Triassic-Jurassic boundary (Silvestri and Szajna, 1993; Szajna and Silvestri, 1996; Szajna and Hartline, 2001; Olsen et al., 2001b). Assemblages in rocks of Triassic age contain abundant *Brachychirotherium* (suchian) and more rare *Apatopus* (phytosaur), a form informally designated “new taxon A” of Szajna and Silvestri (1996) (suchian), as well as *Rhynchosauroides* (lepidosauromorph), *Batrachopus*, (crocodylomorph suchian) a form referred to “new taxon B” (Szajna and Silvestri, 1996) (suchian), and abundant small- to medium-sized dinosaurian tracks usually referred to as various species of *Grallator* and *Anchisauripus* (i.e., *Eubrontes* spp. in the terminology of Olsen et al, 2001b). *Brachychirotherium*, *Apatopus*, and “new taxon B” have never been found in strata of Jurassic age anywhere, despite the global abundance of Early Jurassic age footprint assemblages. The highest footprint assemblage with *Brachychirotherium* and “new taxon B” occurs about 11 m below the “blue-gray sandstone” and the fern spike (Fig. 5). Even closer to the boundary is a poorly sampled footprint-bearing level with *Rhynchosauroides*, *Batrachopus*, and *Grallator* and *Anchisauripus* (*Eubrontes* spp.) that is about 7 m below the “blue-gray sandstone” and fern spike. Although thus far not very productive, only *Grallator* and *Anchisauripus* (*Eubrontes* spp.) have been found above the “blue-gray sandstone” in the Jacksonwald syncline.

Abundant tetrapod bones occur in a zone about 400 m below the Orange Mountain Basalt in the Jacksonwald syncline in member TT (Fig, 3). Thus far, this interval has produced numerous skeletal remains, including skulls and articulated skeletons of the procolophonid parareptile *Hypsognathus fenneri*, and the crocodylomorph cf. *Protosuchus*, as well as other as yet unidentified remains, including probable phytosaur teeth. These bone occurrences are about 800 ky older than the Triassic-Jurassic boundary, and help constrain the ranges of Triassic-type taxa.

Correlation of the Jacksonwald syncline sections with the Newark basin cores is fairly straightforward, despite the muted cyclicity in the uppermost Passaic Formation in the cores. A thin but well-defined interval of reversed polarity (E23r) occurs about 17 m below the Orange Mountain Basalt in the Martinsville no. 1 core (Kent et al., 1995; Kent and Olsen, 1999a). This reversed interval is sandwiched between two very thick normal polarity intervals (E23n and E24n). Magnetic polarity chron E23r has also been identified in the Jacksonwald syncline section (Olsen et al., 1996a). The pattern of prominent gray beds in the Jacksonwald syncline sections and their relationship to chron E23r is matched very closely by the relationship between very thin gray and purple bands and chron E23r in the Martinsville no. 1 core. This laterally repeated pattern permits precise outcrop and core correlation (Fig. 5) (Olsen et al., 1996a), in particular tying the Jacksonwald sections to the stratigraphy in the northern Newark basin where the cyclostratigraphy of the bulk of the Jurassic section has been established by study of the ACE cores (Fig. 5) (Olsen et al., 1996b).

Exposures in quarries and at construction sites in the vicinity of Paterson and Clifton, New Jersey, in the same areas in which the ACE cores were drilled, have produced a series of important fossil assemblages within the upper Passaic Formation. Although a magnetic stratigraphy does not exist for this area of the Newark basin section, the pattern of purple intervals in the ACE cores matches that seen in the Martinsville no. 1 core, allowing lithostratigraphic correlation. Based on this correlation, the Martinsville no. 1 core and the ACE cores appear to have nearly the same accumulation rate for the uppermost Passaic Formation.

The uppermost few meters of the Passaic have produced an enormous number of footprints from a variety of exposures. These footprint assemblages contain only *Rhynchosauroides*, *Batrachopus*, and small to large *Grallator*, *Anchisauripus*, and *Eubrontes* (*Eubrontes* spp.). Critically, these assemblages contain the oldest examples of *Eubrontes giganteus*, a dinosaurian track about 20% larger than any older ichnospecies (Olsen et al., 2001b). At one locality within the footprint-bearing sequence, a gray lens of sandstone and shale has produced a macroflora dominated by the conifers *Brachyphyllum* and *Pagiophyllum*, and the fern *Clathropteris meniscoides*, as well as a poorly preserved palynoflorule dominated by *Corollina* and lacking Triassic-type taxa. This footprint and plant assemblage is thus of earliest Jurassic age, an interpretation supported by its stratigraphic position compared to the Martinsville no. 1 core (Fig. 5).

In close proximity to these footprint and plant localities are the exposures that yielded skeletal remains of *Hypsognathus fenneri*, including the holotype specimen (Sues et al., 2000). These occurrences are about 45 m below the Orange Mountain Basalt. Based on correlation with the Martinsville no. 1 core and the Jacksonwald syncline sections, these represent the youngest examples of typical Triassic osteological taxa in eastern North America. Based on the cyclostratigraphy, the *Hypsognathus*-bearing horizons are within 200 ky of the palynologically defined Triassic-Jurassic boundary.

Above the Orange Mountain Basalt are many horizons in the Feltville, Towaco, and Boonton formations that have yielded very abundant tetrapod footprints of typical Connecticut Valley aspect (Olsen, 1995; Olsen et al., 2001b), as well as several well-preserved *Corollina*-dominated palynoflorules of typical Early Jurassic aspect (Cornet and Traverse, 1975; Cornet, 1977; Cornet and Olsen, 1985).

Thus, the paleontology and Milankovitch cyclostratigraphy of the upper Passaic Formation and succeeding units constrains the Triassic-Jurassic biological transition to within 10 ky based on palynology, and within 20 ky, based on footprints. Because there are no assemblages of bones from Jurassic age strata of the Newark basin, constraints based on osteological taxa are discussed below in a regional context.

Newark Basin Geochemical and Mineralogical Anomalies

The abundances and inter-element ratios of the siderophile elements, such as Cr, Co, Ni, and especially the platinum group elements (PGEs) have been used to investigate the possible presence of a meteoritic component in terrestrial rocks at several geological boundaries (e.g., Alvarez et al., 1980; and references in Montanari and Koeberl, 2000). However, the expected concentrations of PGEs in terrestrial rocks, even with extraterrestrial enrichment is exceedingly low. For example, the addition of about 0.1% of a meteoritic (chondritic) component to a crustal rock would yield an enrichment of about 0.5 ppb Ir to the crustal abundance (~0.02 ppb Ir) in the resulting impact breccia (Koeberl, 1998). Due to these low abundances, only very sensitive analytical techniques, such as iridium coincidence spectrometry (ICS) and inductively coupled plasma source mass spectrometry (ICP-MS), after chemical pre-separation of the PGEs, can be used.

There have been two previous, unsuccessful, attempts to find geochemical and mineralogical anomalies at the Triassic-Jurassic boundary in the Newark basin. Smith et al. (1988) looked specifically for an Ir anomaly in the same units in the Jacksonwald syncline that we examine here, but the amounts present fell below their detectable limits. Mossman et al. (1998) looked for shocked quartz without success in the same interval.

Analytical Methods

We collected the material to be analyzed as channel samples, i.e., contiguous, continuously sampled intervals assuring that no part of the target section could be missed. Because of the high accumulation rate of the sections, channel samples contained about 3 cm of section per sample. However, using this sample strategy, the sacrifice is dilution of any relatively thinner sampled anomaly.

The samples were crushed manually in a plastic wrap, then mechanically in an alumina (ceramic) jaw crusher, and powdered using an automatic agate mill. Major, minor, and trace elements in all samples were analyzed by X-ray fluorescence spectrometry (XRF) and instrumental neutron activation analysis (INAA). Major element analysis was done by standard XRF procedures. The concentrations of the trace elements Sr, Y, Zr, Nb, Ni, Cu, V and Ba were also determined by XRF analysis. For information on procedures, accuracy and precision, see Reimold et al. (1994).

For INAA, which was used to measure the abundances of all other trace elements, aliquots of ~150 mg were sealed into polyethylene (PE) vials. The sample vials were packed with well-characterized reference material into a larger PE-irradiation vial. The international granite standards AC-E and U.S.G.S. G-2 (Govindaraju, 1989), the Allende meteorite standard reference powder (Jarosewich et al., 1987), and - for Au and Ir - the mineralized gabbro PGE standard WMG-1 (CANMET, 1994) were used as certified reference materials to determine the accuracy of the analysis.

The samples were irradiated at the TRIGA Mark II type reactor at the Atominstitut der österreichischen Universitäten in Vienna for 7 hours at a flux of about 2×10^{12} $ncm^{-2}s^{-1}$ (Koeberl, 1993).

Due to the low Ir abundances encountered during INAA, the Ir content was additionally determined with the ICS system at the Institute of Geochemistry at the University of Vienna. Seven crushed and powdered samples of about 50 mg each, as well as a series of the Allende meteorite reference sample (Jarosewich et al., 1987) that were diluted with high purity quartz powder to produce standards with Ir concentrations between 35 ppt and 6.93 ppb, were sealed into high purity quartz glass tubes. Samples and standards were packed into aluminum foil and an aluminum capsule, and irradiated for 24 to 48 at a flux of about 7 x 10^{13} $ncm^{-2}s^{-1}$. After a cooling period of about ten weeks, the samples were first measured for five to eight hours. The lines of ^{192}Ir at 316 and 468 keV were used, and the method requires that only coincident signals at both lines are used for further processing. Samples that yielded results close to the detection limit were measured for at least another 24 hours. Regression analysis was done for the dilution series to provide good precision and accuracy. The calculated peak volumes were used to perform the unknown sample analysis with live time correction, decay time correction, flux correction, background subtraction and comparison with the standards. The precision of the Ir measurements follows a logarithmic error function with the lowest relative errors in the highest concentrations (e. g., 21±9 ppt vs. 285±33 ppt). For further details on this method (precision, accuracy), see Koeberl and Huber (2000).

Results and Discussion

Undeterred by the previous unsuccessful attempts, we examined Ir and other elemental concentrations at four sections along strike in the Jacksonwald syncline directly around the fern spike (Fig. 8). The samples show variations in Ir content from 19 to 285 ppt (0.2 - 2.9 ng/g) (Table 1). All sections except Section I show a distinct Ir anomaly directly at the boundary with a distinct systematic association between Ir content and stratigraphy. The elevated levels of Ir are mostly associated with higher levels of Al in a white smectitic claystone (Smith et al., 1988), directly beneath the thin coaly layer (Figs. 6, 7), although there is no correlation between Al and Ir in the data in general. The anomaly is directly associated with the previously identified fern spike in these sections, recalling the similar pattern at the Cretaceous-Tertiary boundary in the western US (Tschudy et al., 1984; Nichols and Fleming, 1990). It is possible that the relatively weak Ir anomaly (relative to the apparent background) seen thus far is a consequence of dilution by the rather coarse sampling level (ca. 3 cm per sample) required by the very high accumulation rates (~1 m / 2000 yr) in the sampled part of the Newark basin. We can probably rule out a simple diagenetic concentration of Ir along a redox boundary because of the good correlation between Ir and stratigraphy, despite the lateral facies change from gray and black strata in the east to virtually entirely red strata in the westernmost section (Grist Mills, Fig, 8). Indeed, the sample with the highest Ir content at the Grist Mills section is red (177 ppt. sample TJ 28).

In the one section (I) that did not show a systematic association between Ir content and stratigraphy, Ir levels were highest in the "blue-gray sandstone" rather than in the spore-rich clays just below it. This could be because our sample of the sandstone (sample TJ 1) contained mud chips eroded from a presumably Ir-enriched layer upstream. The sandstone was bulk processed and it will be necessary to run separate analyses on the sandstone and clay pebble separates to test this hypothesis.

The situation does not, unfortunately, become any clearer when considering the concentrations of other elements. These data are discussed here in only a perfunctory and very preliminary way, recognizing that much more might be done with the data. In addition to measuring Ir by ICS, we determined the abundances of a total of 44 major and trace elements in all samples (Table 1). Comparing the trends of the Ir data with those of other siderophile elements - such as Co, Ni, and Cr, which are often used as tracers of meteoritic components - does not yield distinct correlations. For example, in section I, the highest Ni and Co abundances are at samples TJ 2 and TJ 3, whereas Ir shows more variation in the section. No correlation is apparent for samples of section II and the variations in section III do not appear to be significant. On the other hand, in section IV, those samples with the highest Ir contents (TJ 28 – 30) do not have significantly elevated Cr, Co, and Ni abundances. However, Ni does show an increase from about 20 ppm to about 50 ppm over the Ir maximum, but marking out a broader peak. The values for Co show more variation, but follow the same trend as for Ir and Ni.

The trace element data do not seem to support the idea that the Ir enrichments could be related to volcanic ash. Elements that might correlate in abundance with Ir if the source were altered volcanic material (e.g., Cs, Al, Cu, and V) , reveal no significant correlations (Table 1). The variations in these elements in section I are irregular. In section IV, the concentrations of Cs vary with no correlation to the distinct Ir enrichment and associated Co and Ni abundance peaks. In fact, the highest Cs abundances occur for samples outside of the zone of siderophile element enrichment. Thus, we cautiously consider a volcanic interpretation of the Ir enrichments unlikely, although we refrain from more detailed comment on the trace element data until more thorough sampling, especially further away from the palynologically identified boundary, can be performed.

A significant problem with our analysis of the association between Ir abundance and the fern spike in the Jacksonwald syncline sections is that the geochemical and palynological analyses were not conducted on the same samples, or even at the same sections, and there is considerable variability from section to section in both variables. In an attempt to mitigate this problem and facilitate comparison of the Ir and pollen and spore data, we have averaged the Ir data and combined the spore data from all of the Jacksonwald synclines sections examined to date, using the base of the "blue-gray sandstone" as the correlative datum (Figs. 7, 9). No attempt has been made to account for possible lateral changes in accumulation rate.

The averaged and combined data show a strong correlation between spore percentage and Ir content. However the spore maximum is below the Ir maximum and the spore "spike" appears quite broad. This could reflect an actual offset between the two data sets, or alternatively, it could reflect small variations in either accumulation rate or depth of erosion of the overlying "blue-gray sandstone" at different sections. Existing data do not permit these hypotheses to be tested. Clearly, what is needed is very detailed sampling of sections, in which splits of the same samples are subjected to palynological and geochemical analyses. In addition, Ir measurements are clearly needed from a broader stratigraphic swath around the boundary to assess background Ir levels. The available data are, however, very encouraging and do suggest that there is a modest Ir anomaly at the biologically-identified boundary and that it is associated with a fern spike.

Twenty grams of samples TJ 3-6 , 9, 10, 17, 18, 20, 21, and 22 were broken down in water and sieved for shocked quartz, and about 100 to 300 grains of the 88-149 _m mineral fraction were examined under plain and polarized light. All quartz grains observed are angular, and few are clear. Most of them contain mineral and fluid inclusions. In some cases, fluid inclusions are aligned and in a very few cases they seem to be aligned along parallel lines, but none of these features are characteristic for shock metamorphism. Besides these very few cases, no planar features (PFs) or lamellae potentially representing planer deformation features (PDFs) were observed. Thus, as was the case for Mossman et al. (1998), our search for shocked quartz grains was not successful.

The Triassic-Jurassic Boundary in Other Pangean Continental Rifts

The patterns seen in boundary sections in other continental rift basins in the Newark Supergroup and Morocco are consistent with that seen in the Newark basin, but also provide important additional information. The most southern rift known to preserve the Triassic-Jurassic boundary is the Culpeper basin of Virginia. The boundary section there has been described by Fowell (1994), but currently known outcrops and exposures have not permitted detailed stratigraphic or palynological analysis.

The next basin northward, for which there is significant stratigraphic information relevant to the Triassic-Jurassic boundary is the Hartford basin of Connecticut and Massachusetts (Fig. 2). A section exposed in the Cinque Quarry in East Haven, Connecticut, reveals the uppermost New Haven Formation and overlying pillowed base of the Talcott basalt (Heilman, 1987). Nearly the entire New Haven Formation is fluvial, composed of brown and minor gray coarse clastic rocks and red mudstone. The uppermost 40 cm of the New Haven Formation at the Cinque Quarry is, in contrast, gray, and may represent marginal lacustrine environments (Fig. 7). The uppermost few centimeters of gray mudstone and sandstone preserve abundant *Brachyphyllum* shoots and cones and a palynoflorule of typical Early Jurassic aspect, dominated by *Corollina* (Robbins, quoted in Heilman, 1987). Because there is no dispute that most of the New Haven Formation is of Late Triassic age (Cornet, 1977), the Triassic-Jurassic boundary probably lies either within the gray sequence below the conifer bearing level, or closely underlying it in the red beds.

Strata interbedded with and overlying the basalts of the Hartford basin preserve a cyclostratigraphy nearly identical to that of the Newark basin, implying nearly exact synchrony of climatic and eruptive events (Olsen et al., 1996b). However, compared to the youngest formation in the Newark basin (Boonton), the Portland Formation which overlies the youngest basalt formation in the Hartford basin (Hampden Basalt), is much thicker (~4 km) and represents a much longer time than is preserved in the Newark basin. Hence, the Portland Formation provides an important supplement to the Newark basin astronomically calibrated time scale and allows the floral and faunal change seen at the Triassic-Jurassic boundary to be placed into a more extensive temporal perspective. The lower Portland Formation exhibits a cyclostratigraphy virtually identical to that of the Boonton Formation of the Newark basin. However, while less than 300 ky of Boonton Formation was recovered by the ACE cores, at least 3 my is represented by the Portland Formation. Recently, the magnetostratigraphy of roughly 2 my of the lower parts of the Portland has been determined from outcrop samples (Kent and Olsen, 1999b). Although clearly not known at the same level of detail as the Newark basin time scale, which is based on continuous core, the Portland Formation magnetostratigraphy does constrain the maximum duration of chron E24n, the top of which is not seen in the Newark basin record (Kent and Olsen, 1999a) (Fig. 4). The polarity reversal stratigraphy of the Portland Formation also provides a critical link to the marine polarity sequence from the Paris basin (Yang et al., 1996), indicating that the upper part of the Portland Formation is of Sinemurian age. Based on the Newark basin time scale, with the addition of the Portland Formation information, the duration of the Hettangian is thus about 2 my, in excellent agreement with new information from U-Pb dates from marine sections in British Columbia (Palfy et al., 2000b).

Palynoflorules from the Jurassic part of the Hartford basin section show only minor changes through time (Cornet, 1977; Cornet and Olsen, 1985), predominately involving the appearance of new taxa and species-level changes in the dominant pollen taxon *Corollina*. Vertebrate footprint and bone assemblages from Hettangian and Sinemurian age strata of the Hartford basin show no obvious changes at all, except perhaps an increase in the size of some footprint forms (e.g., *Anomoepus*: Olsen and Rainforth, 2001). The slow change through the Hettangian and into the Sinemurian, spanning at least 3 my, contrasts dramatically with the extraordinarily abrupt change seen at the Triassic-Jurassic boundary, which conservatively took less than 20 ky.

In predrift configuration, the next continental rift basins north for which there is significant information on the Triassic-Jurassic boundary are those in Morocco, particularly the Argana basin (Fig. 2). The stratigraphy of the boundary section in the Argana basin differs from that in the Newark basin in that cyclical gray and black strata are limited in outcrop to a couple of meters below the Argana basalt (Olsen et al., 2000). Palynoflorules from these very thin (<20 cm) gray and black mudstones, which are interbedded with red mudstones, show a transition from an assemblage with *Patinasporites* to one without this taxon or any other forms typical of the Triassic. Instead it is dominated by *Corollina* (Olsen et al., 2000). The physical stratigraphy of this sequence, despite its very condensed appearance, as well as that of the overlying Argana Basalt and Amsekroud Formation, is closely comparable to that of the Newark basin section, implying a nearly identical sequence and timing of events (Olsen et al., 2001a).

Of special interest is the physical stratigraphy of the Khémisset basin in Morocco. In outcrop, the stratigraphy of the sedimentary section immediately below the oldest basalt is virtually identical to that seen in the Argana basin. However, in the subsurface, these cyclical red, gray and black mudstones pass into interbedded black, red, and white halite and potash salts. Our preliminary palynological results from outcrop sections suggest that the Triassic-Jurassic boundary lies in the cyclical mudstone sequence just below the oldest basalt, and that the boundary should be present within the salt in the subsurface as well – a hypothesis we are presently examining.

Where studied so far, the physical stratigraphy of the sequence of basalts and the directly underlying and overlying sediments of the other Triassic-Jurassic which crop out in basins of Morocco is very similar to that in the Newark and Argana basins, albeit very condensed. Thus, we argue for correlation of biotic and tectonic events at a very fine scale across the Moroccan basins. We hypothesize that the similarity, in detail, of the sequences just below the oldest basalts in all these basins to that seen in the Argana and Newark basins demarcates the position of the Triassic-Jurassic boundary. However, published interpretations of the biostratigraphy of these basins differs dramatically from that presented here (Figs. 6, 7). According to the summary of Oujidi et al. (2000) the ages of these homotaxial sequences are dramatically different in different basins, ranging in age from Ladinian to Norian. We attribute these differences in interpretation to a lack of appreciation of floral provinciality, the importance of the fact that earliest Jurassic assemblages are characterized by survivor assemblages rather than the appearance of new taxa, and by reliance on ostracode and bivalve taxa of dubious identification and biostratigraphic utility. Palynological work by us presently underway on various sections in Morocco will test our hypothesis.

The Fundy basin of the Maritime provinces of Canada is the next outcropping continental rift basin to the north, and the basin for which there is the most information on the Triassic-Jurassic boundary after the Newark basin. Fowell and others (Fowell and Olsen, 1993; Fowell, 1994; Fowell et al., 1994) have shown that the Triassic-Jurassic boundary is preserved within the uppermost few meters of the Blomidon Formation, below the North Mountain Basalt (Fig. 6) (Kent and Olsen, 2000b). The boundary section is very condensed and closely comparable to that in the Argana basin (Fig. 6). Like the Newark basin, the palynology of the boundary is marked by an abrupt disappearance of assemblages with *Patinasporites* and their replacement by assemblages dominated by *Corollina*. Although no fern spike has been found in the section studied in detail by Fowell and Traverse (1995) at Partridge Island (Cumberland County, Nova Scotia), outcrops of the uppermost few meters of Blomidon Formation at Central Clarence (Annapolis County, Nova Scotia), have one layer that produces a macroflora consisting entirely of the fern *Cladophlebis* (Carroll et al., 1972) (Fig. 6). This is the only foliage macroflora from the Fundy basin of Nova Scotia, and the fact that it consists entirely of ferns, at the expected position of the Triassic-Jurassic boundary, suggests to us that it may in fact represent the fern spike itself. Unfortunately, the exposures at Central Clarence preclude a detailed study of this section without significant excavation.

The cyclostratigraphy of the lower McCoy Brook Formation, overlying the North Mountain Basalt, is closely comparable to the that of the Feltville Formation of the Newark basin and its equivalents in other rifts in the eastern United States (Olsen et al., 1996b; 2001a). A rich vertebrate assemblage has been recovered from the lower McCoy Brook Formation and is notable for the abundance of well-preserved tetrapod bones and skeletons (Olsen et al., 1987). This assemblage is the oldest known from the continental Early Jurassic that is reasonably diverse and well dated. Despite the sampling of both terrestrial and aquatic habitats, typical Triassic osseous and footprint taxa are completely absent. In addition, while the remains of small tetrapods are very common at multiple levels, procolophonids are absent. Thus, this assemblage helps constrain the osseous record of at least some of the tetrapod extinctions to within 300 ky around the boundary. Footprint assemblages from the rest of the outcropping McCoy Brook Formation are consistent with this picture and are entirely of Connecticut Valley aspect (i.e., Early Jurassic).

Anders and Asaro (reported in Olsen et al., 1990) examined the upper 100 meters of the Blomidon Formation for shocked quartz and Ir anomalies but were unsuccessful. Mossman et al. (1998) examined the stratigraphic region near the Triassic-Jurassic boundary, including at Partridge Island, for shocked quartz and Ir. Although some planar features were seen, they concluded that none were characteristic of shocked metamorphism. They also concluded that although no distinct Ir anomaly was found, the highest Ir amounts were found in proximity to the Triassic-Jurassic boundary. We stress that these results are not inconsistent with our results from the Newark basin, but that much tighter geochemical sampling directly tied to the stratigraphic interval with the biotic turnover needs to be conducted in the Fundy basin.

Thus, the emerging picture of faunal and floral change around the Triassic-Jurassic boundary in eastern North America and Morocco is one of extraordinarily rapid, synchronous change over a very large area. This change, at least in the Newark basin and possibly in the Fundy basin, is associated with a geologically abrupt burst in fern abundance suggestive of major ecological disruption. It is important to note that within this context, with the exception of the appearance of the large dinosaurian ichnospecies *Eubrontes giganteus*, the earliest Jurassic floral and tetrapod assemblages consist entirely of survivor taxa with no originations and no apparent replacement of Triassic forms by Jurassic ecological vicars. Indeed, it has been hypothesized that even the appearance of *Eubrontes giganteus* may represent a consequence of ecological release upon the extinction of Triassic competitor forms, largely members of the non-dinosaurian Crurotarsi, such as rauisuchians and phytosaurs (Olsen et al., 2001b). It is also important to note that the Triassic-Jurassic boundary marks the end of the persistent floral and faunal provinciality that characterized the Triassic, and the establishment of a nearly cosmopolitan terrestrial community (Cornet and Olsen, 1985; Olsen and Galton, 1984; Sues et al., 1994). The biological data available thus far are consistent with a catastrophic end to the Triassic comparable in magnitude and similar in pattern to that characterizing the terminal Cretaceous event. Our new results from the Newark basin Triassic-Jurassic boundary do show a modest Ir anomaly associated with the biotic turnover and the fern spike, that again is remarkably similar to that described for the Cretaceous-Tertiary boundary. We stress that these results do need to be tested by much more widespread geochemical and palynological analyses, both stratigraphically and geographically.

Relationship to Possible Impact

It is clear that, at least superficially, there is a strong similarity between the Cretaceous-Tertiary boundary in the western North American interior and the Triassic-Jurassic boundary in eastern North America. This similarity includes the specific pattern of floral and faunal extinction, the fern spike and now the presence of an apparent Ir anomaly. The similar patterns might indicate similar cause, and an impact origin for both boundaries has been suggested (Dietz, 1986; Olsen et al., 1987, 1990). Although shocked quartz has not been reported from eastern North America, it has been reported by Bice et al. (1992) from a Triassic-Jurassic boundary section in Tuscany, and there is an additional report from the Kendelbach section in Austria (Badjukov et al., 1987). However, in both cases the shocked quartz was identified only petrographically, which is now not considered definitive (e.g., Grieve et al., 1996; Mossman et al, 1998), and in neither case has there been a subsequent attempt at independent confirmation. Of course, if the impact site were very distant, or in oceanic crust, it is possible that shocked quartz would be very rare or absent.

Originally, the giant Manicouagan impact was suggested as a possible cause (Olsen et al., 1987) of the Triassic-Jurassic mass extinctions. However, U-Pb dates from this feature by Hodych and Dunning (1992) suggest that its age is 214±1 Ma, which is consistent with older $^{40}Ar/^{39}Ar$ and Rb-Sr dates from the impact, but incompatible with the dates of the basalts overlying the boundary (~200 Ma), which at least at the level of uncertainty of radiometric dates, should be the age of the boundary.

Relationship with CAMP

A remarkable aspect of the Triassic-Jurassic boundary is the very close proximity in both stratigraphic thickness and time (~20 ky) to the oldest exposed CAMP flood basalts in eastern North America and Morocco. The CAMP tholeiites may represent the largest known (in area at least) igneous event in Earth history, covering an area of 7×10^6 km^2 (e.g., Marzoli et al., 1999) (Fig. 2). The pre-erosion volume of the CAMP may have been in excess of 3×10^6 km^3, making it larger than any other known continental flood basalt province. The close association between the CAMP lavas and the boundary has led to speculation that the extinctions might have been caused by climatic changes resulting from gas and aerosol emissions from the eruptions (Courtillot et al., 1994; McHone, 1996; Marzoli et al., 1999; Palfy et al., 1999). McElwain et al. (1999) found that stomatal density in a range of plant taxa drops significantly within the same taxa at the florally-identified Triassic-Jurassic boundary in Greenland (Kap Stewart Formation). This change is direct evidence suggesting a major increase in CO_2 at the boundary which they speculate might be due to CAMP volcanism. Smith et al. (1988) noted that a volcanic source could be responsible for the smectitic clay at the boundary in the Jacksonwald syncline of the Newark basin. It is also possible that an Ir anomaly, especially a modest one, could be explained by deep-seated basaltic volcanism as suggested by Olmez et al. (1986) for the Cretaceous-Tertiary boundary, although this hypothesis is not supported by our geochemical analyses of associated elemental concentrations.

A very significant problem with the volcanic hypothesis for the origin of the Triassic-Jurassic mass extinction is that wherever the biological signature of the mass extinction and the oldest CAMP basalts have been observed in the same section, the basalts invariably postdate the extinctions (as described above), albeit by only a short time. However, the basalts that are known in superposition with the boundary amount to a small part of the CAMP; their temporal relationship with the rest of the CAMP and the boundary is unknown at the required fine level of resolution.

Olsen (1999) and Olsen et al. (2001a) point out that there is some slim paleomagnetic evidence for CAMP eruption plausibly occurring just prior to the boundary consisting of very rare dikes of CAMP radiometric age (i.e., 200 Ma) with reversed magnetic polarity. Because all the CAMP basalts above the boundary are uniformly of normal polarity, this shows that SOME of the CAMP igneous activity occurred at a different time than the known flows. The temporally closest interval of reversed polarity is E23r, located just below the boundary (Figs. 4-7). If further research confirms the reality of the reversed dikes, a significant portion of the CAMP could easily be pre- or syn-boundary. For example on huge basaltic edifice that could predate the boundary includes the massive seaward-dipping reflectors off the southeastern United States that may be part of CAMP (Holbrook and Kelemen, 1993; Olsen, 1999; Olsen et al., 2001a) and could be volumetrically as large as the rest of the province. Thus we do not reject the possibility that at least part of the CAMP, perhaps even the largest part, could have been emplaced just before or during the Triassic-Jurassic mass extinctions.

Impact AND Volcanism?

Rampino and Stothers (1988) and Courtillot et al. (1994) have shown that based on a compilation of published radiometric dates of igneous rocks and literature ages for geologic boundaries, there is a very good correlation between major continental flood basalts and mass extinctions over the last 300 my. Particularly prominent are the three largest Phanerozoic mass extinctions (Permo-Triassic, Triassic-Jurassic, and Cretaceous-Tertiary) and their remarkably tight association with the three largest Phanerozoic continental flood basalts (Siberian, CAMP, and Deccan, respectively). It seems to us very hard to dismiss this correlation as a coincidence and it is particularly interesting that at least two (Triassic-Jurassic and Cretaceous-Tertiary) have evidence of an impact (the latter impact evidence is unimpeachable). Thus, we cannot dismiss the very speculative possibility that perhaps the flood basalt volcanism was somehow triggered or enhanced by an impact (e.g., Boslough et al., 1996), despite the fact that preliminary models of the energetics of impacts suggest that causing volcanism *de novo* with an impact would be very difficult to do (e.g., Melosh, 2000).

Conclusions

The biotic pattern around the continental Triassic-Jurassic boundary has many similarities with the much better understood Cretaceous-Tertiary boundary, including the very short duration of the extinction event, its selectivity, the composition of post-boundary assemblages made up of only survivor taxa, and the presence of a regional "fern spike" at the microfloral extinction level. We have shown here new evidence of at least a modest iridium anomaly associated with the fern spike. A summary of the data for the Triassic-Jurassic boundary based mostly on the Newark basin is shown in Fig. 10. We have also shown the enormous area over which the stratigraphy of the continental boundary is remarkably consistent. Also shared with both the Cretaceous-Tertiary and Permo-Triassic boundaries are temporally associated massive continental flood basalts. Unlike the Cretaceous-Tertiary boundary, though, evidence for an impact at the Triassic-Jurassic boundary is not yet conclusive. We also regard it as difficult to dismiss as coincidental the co-occurrence of mass extinctions and flood basalts, including at the Triassic-Jurassic boundary. The newly reported Ir anomaly could be consistent with either an impact or deep-seated volcanic origin, although the latter receives no support from the trace element concentrations or stratigraphic relationships reported here. The microstratigraphy is very similar to continental Cretaceous-Tertiary boundary sections, and this lithological similarity is matched by a similar biotic pattern. However, without additional, more stratigraphically extensive sampling, we cannot completely rule out a volcanic or other non-impact origin for this anomaly.

Acknowledgments.

Work on this project by Olsen, Et-Touhami, and Kent was funded by grants from the US National Science Foundation to Olsen and Kent (EAR-98-14475 and EAR-9804851), and grant from the Lamont Climate Center to Olsen. Work in Morocco was aided by logistical support from ONAREP, for which we are very grateful. Et-Touhami was supported during work on this project by a fellowship from the Fulbright Foreign Student Program (AMIDEAST). The lab work in Austria was supported by Austrian Science Foundation grant Y58-GEO (to Koeberl). Access to the fossil collections of the Nova Scotia Provincial Museum in Halifax was provided by Robert Grantham, who we thank. The paper was reviewed by Emma Rainforth prior to submittal, and David Mossman and Roy Schlische after submittal, which greatly improved it. This is LDEO contribution 6155.

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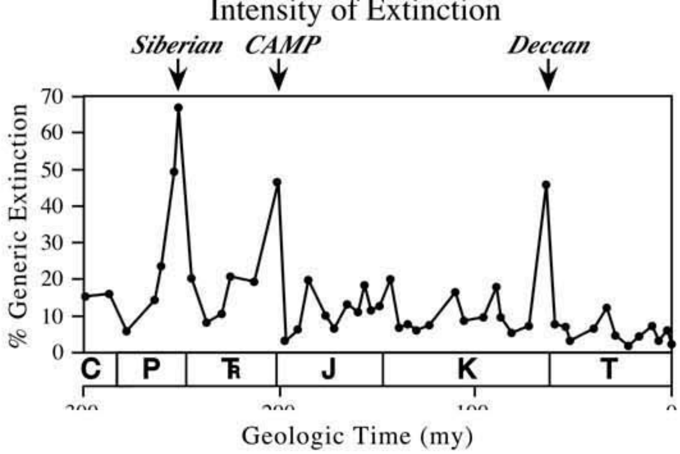

Figure 1. Generic-level extinctions of marine, shelly organisms during the last 300 my, and the distribution of giant flood basalt provinces (in italics). Modified from Sepkoski (1997) to reflect ou estimate of the age of the Triassic-Jurassic boundary.

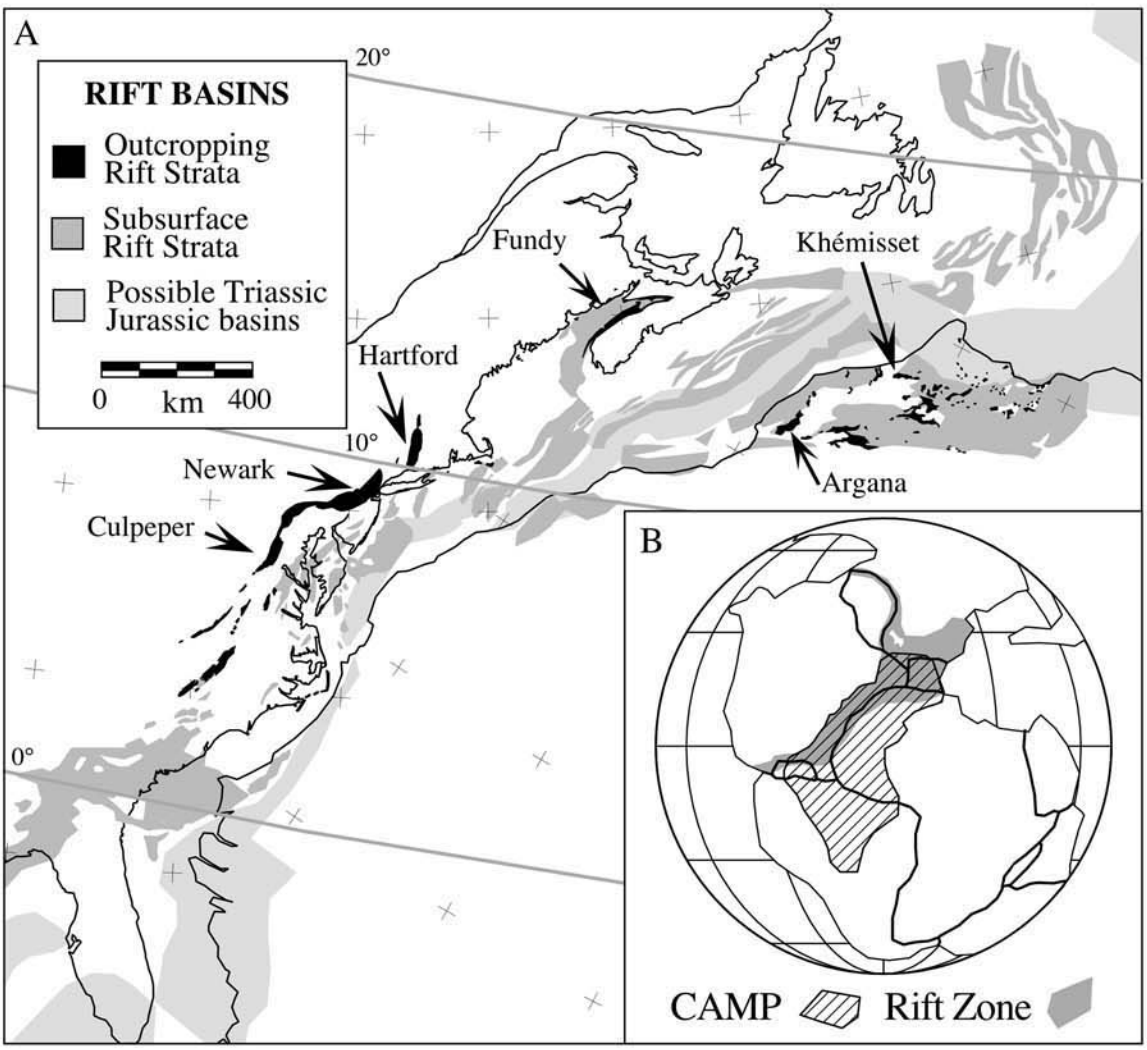

Figure 2. Distribution of central Atlantic margin rift basins of central Pangea and the temporal distribution of the Central Atlantic Magmatic province (CAMP). A, Rift basins of the central Atlantic margins of North America and western Africa in predrift coordinates (for the Late Triassic) showing basins discussed in this paper (Modified from Olsen, 1997); B, Pangea during the earliest Jurassic showing the distribution of the CAMP, overlapping much of the Triassic-Jurassic rift zone (based on Olsen, 1999).

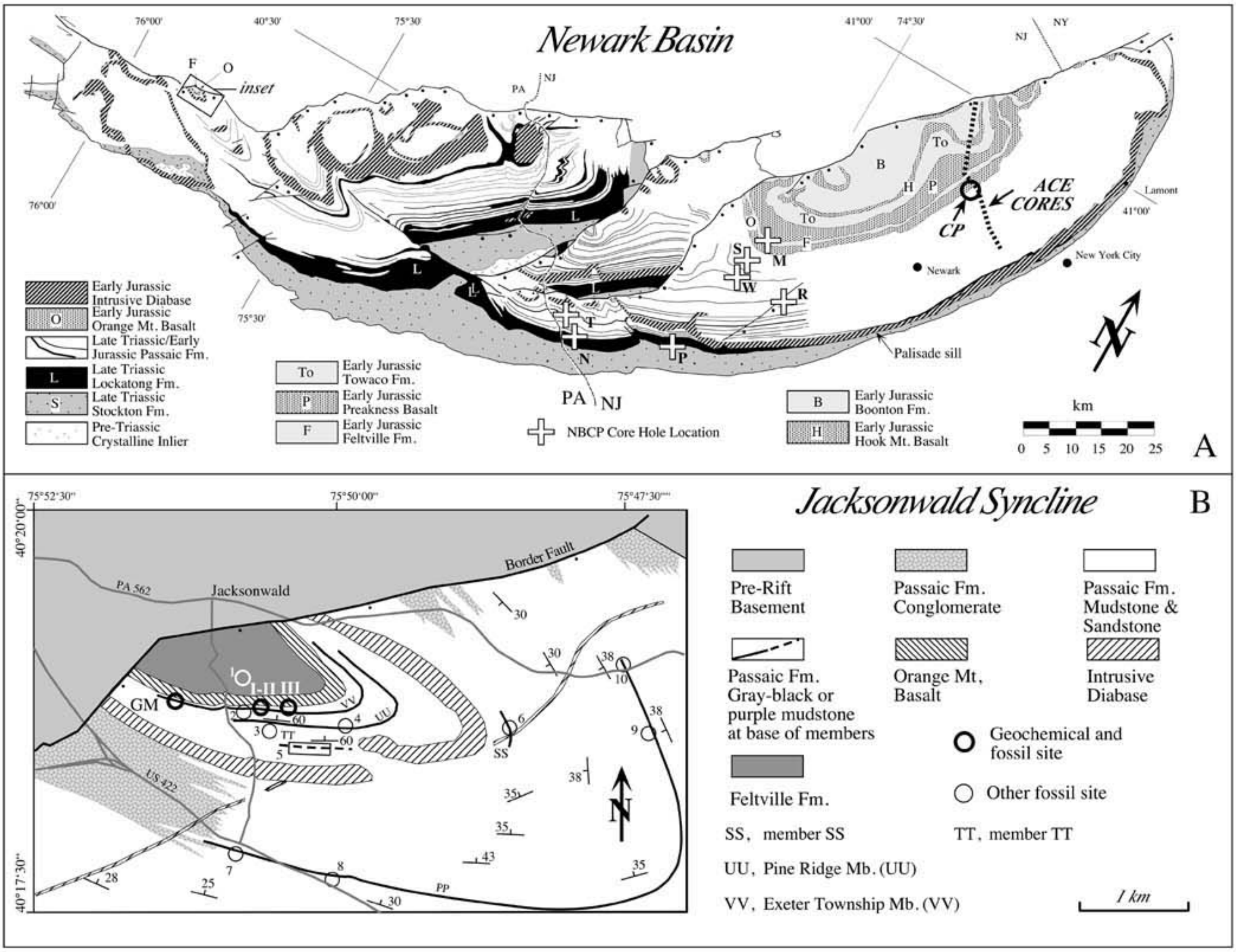

Figure 3. Position of key sections and cores within the Newark basin of New Jersey, New York and Pennsylvania. A, The Newark basin. Letters in bold identify the cores of the Newark Basin Coring Project (NBCP): M, Martinsville; N, Nursery; P, Princeton; R, Rutgers; S, Somerset; T, Titusville; W, Weston. CP, indicates the Clifton-Paterson area and the area in the box outlines the Jacksonwald syncline map of B (below). B, Map of the Jacksonwald syncline showing the positions of the sections discussed in text. Locations for the four sections shown in figure 8 are as follows: GM, Grist Mills (Late 40°18.85 Long. 075°51.20); I, Section I (Lat. 40°18.76 Long. 075°50.56); II, Section II (Late 40°18.76 Long. 075°50.55); III, Section III (Late 40°18.81 Long. 075°50.38). Other paleontological localities are: 1, Exeter Golf Course Estates (Feltville Fm. locality for Eubrontes giganteus); 2, original palynological boundary sections of Cornet (1977) and Fowell (1994); 3, Wingspread footprint locality of Silvestri and Szajna (1996); 4, "Pine Ridge creek" locality for pollen and footprints; 5, Pathfinder Village bone assemblage in member TT (discovery site at (Late 40°18.55 Long. 075°50.10); 6, Walnut Road phytosaur tooth locality in member SS; 7, Shelbourne Square (Ames) footprint locality of Szajna and Silvestri (1993) and Silvestri and Szajna (1993); 8, Heisters Creek development footprint locality; 9, Tuplehocken Road footprint locality; 10, pollen localities OLA1 and OLA3 of Cornet (1977). Specific latitude and longitude coordinates not given here are listed in Olsen et al. (2001b).

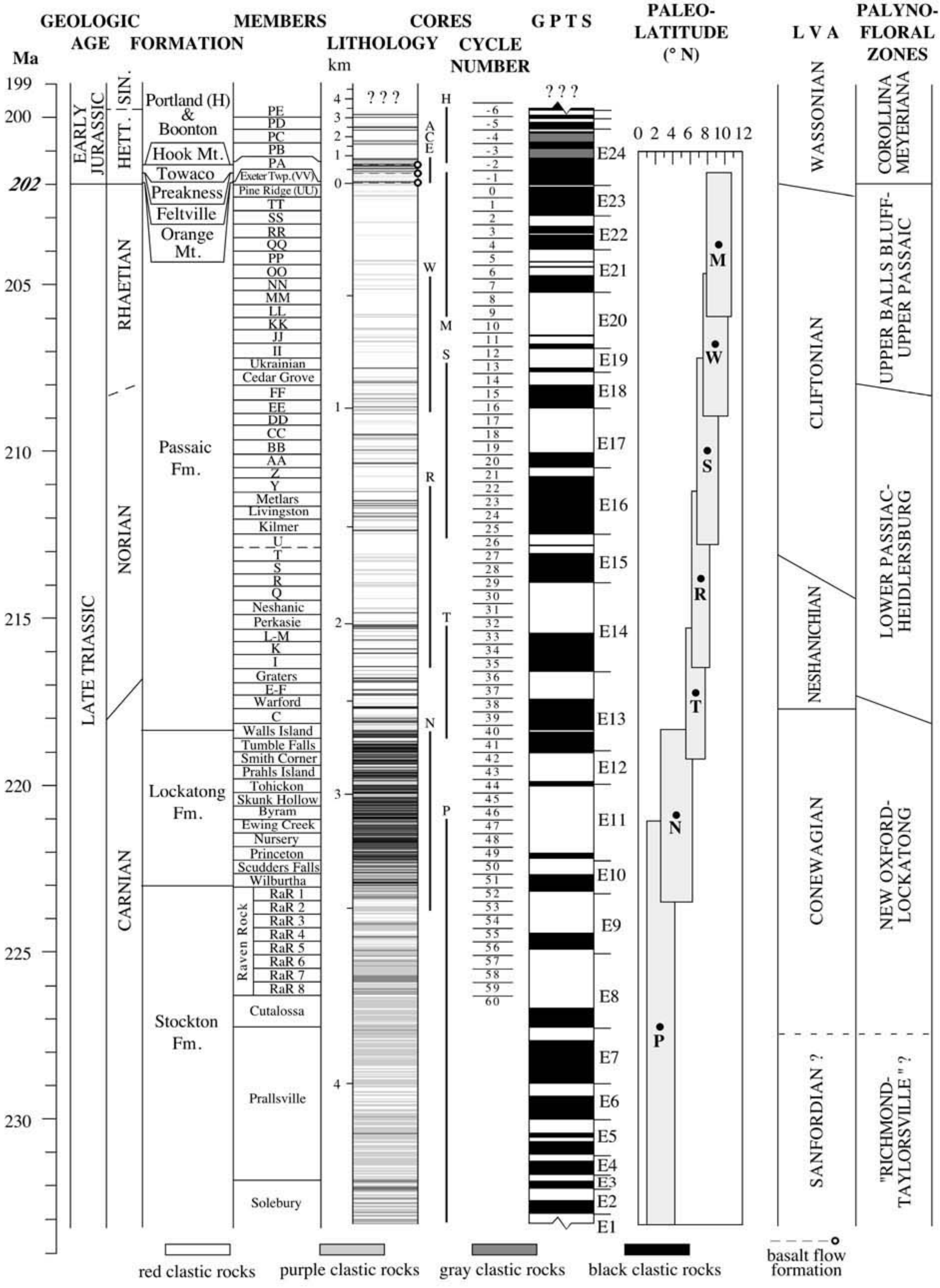

Figure 4. Time scale for the Late Triassic and Early Jurassic based on geomagnetic polarity time scale (GPTS) and astronomical calibration from the Newark Basin Coring Project (Kent and Olsen, 1999a; Olsen and Kent, 1999), the ACE cores (Olsen et al., 1996), and preliminary results from the Hartford basin (Kent and Olsen, 1999b). Biostratigraphic data from Huber et al. (1996), Lucas and Huber (2001), Cornet (1977), and Cornet and Olsen (1985). For the GPTS, black is normal polarity, white is reversed polarity, and gray represents intervals for which there is incomplete sampling (Hartford basin section only). Abbreviations are: ACE, ACE (Army Corps of Engineers) cores; H, Hartford basin section; HETT., Hettangian; L.V.A., Land Mammal Ages; M, Martinsville (NBCP) core; N, Nursery (NBCP) core; P, Princeton (NBCP) cores; R, Rutgers (NBCP) cores; S, Somerset (NBCP) cores; SIN., Sinemurian; T, Titusville (NBCP) core; W, Weston Canal (NBCP)cores. Cycle number refers to the 404 ky cycle of eccentricity with lines placed at the calculated minima.

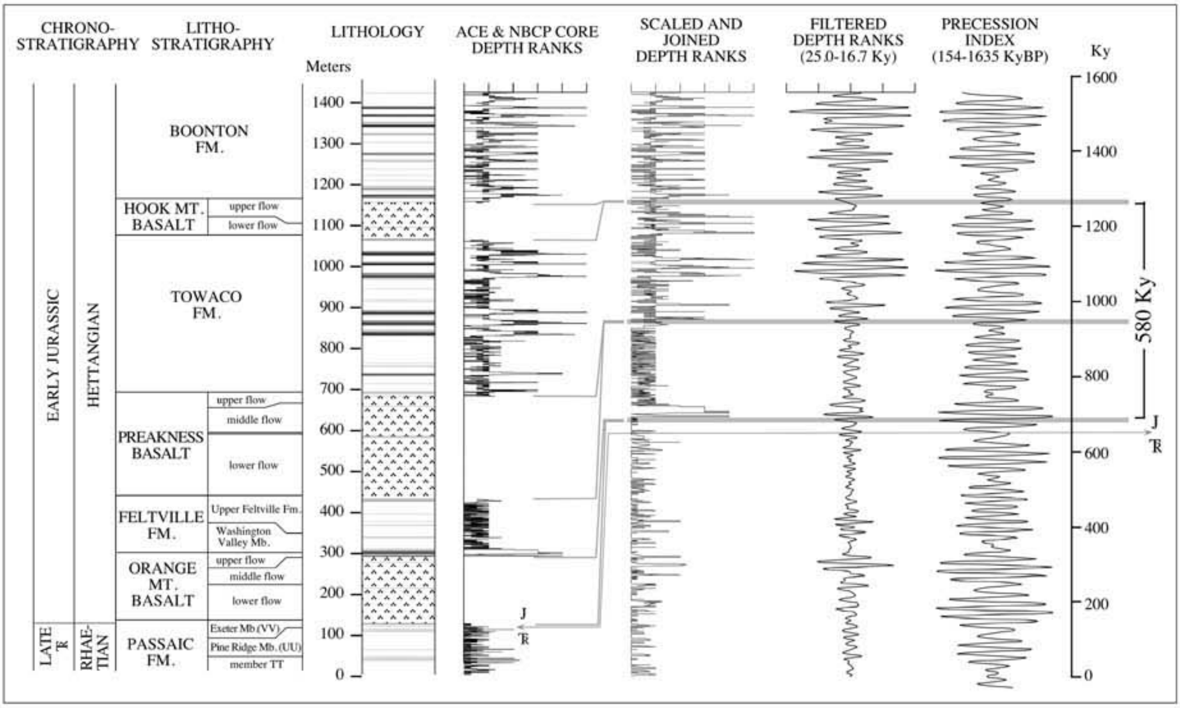

Figure 5. Cyclostratigraphic calibration of the Triassic-Jurassic boundary and succeeding extrusive zone flows and interbedded and overlying sedimentary strata (adapted from Olsen et al., 1996b). Depth ranks are a numerical classification of sedimentary facies sequences in order of increasing interpreted relative water depth (Olsen and Kent, 1996). Comparison of the depth rank curves with an arbitrary segment of a precession index curve indicates that it is not necessary to assume any significant time is represented by the lava flow formations themselves and that the entire flow sequence was probably deposited during an interval of less than 600 ky. Note also that the Triassic-Jurassic boundary (correlated to the Jacksonwald syncline by magneto- and lithostratigraphy - Fig. 7) lies about 20 ky below the Orange Mountain basalt. The depth rank record from the strata above the Preakness Basalt is based on the ACE cores, whereas that from the Passaic and Feltville Formations is based on the Martinsville no. 1 core of the NBCP.

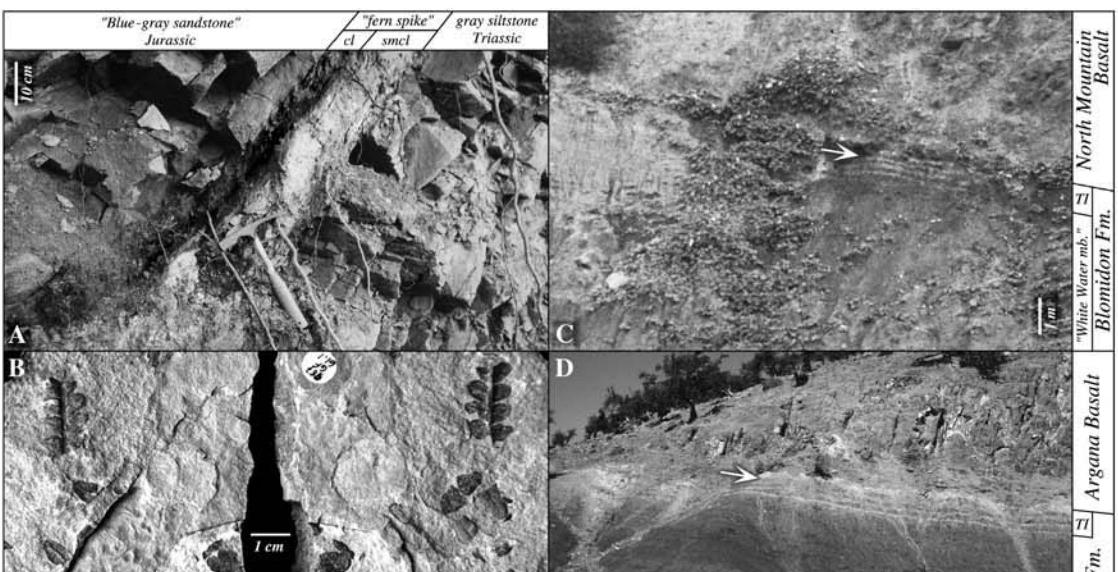

Figure 6. Photographs of sections and ferns of the Triassic-Jurassic Boundary. A, Triassic-Jurassic boundary at section I (Figs. 2, 7): cl, coal and carbonaceous shale; smcl, smectitic claystone. Note that strata dip 60° to north (left). B, part and counterpart of slab bearing fronds of the fern Cladophlebis from the Triassic-Jurassic boundary of the Fundy basin at Central Clarence (Nova Scotia Provincial Museum no. 982.GF.G1.1). C, Triassic-Jurassic boundary at Partridge Island, Nova Scotia (studied by Fowell and Traverse, 1995); TI indicates palynological transition interval and arrow shows position of palynological Triassic-Jurassic boundary. D, Triassic-Jurassic boundary near Argana, Morocco in the Argana basin (Olsen et al., 2000); TI indicates palynological transition interval and arrow shows position of palynological Triassic-Jurassic boundary

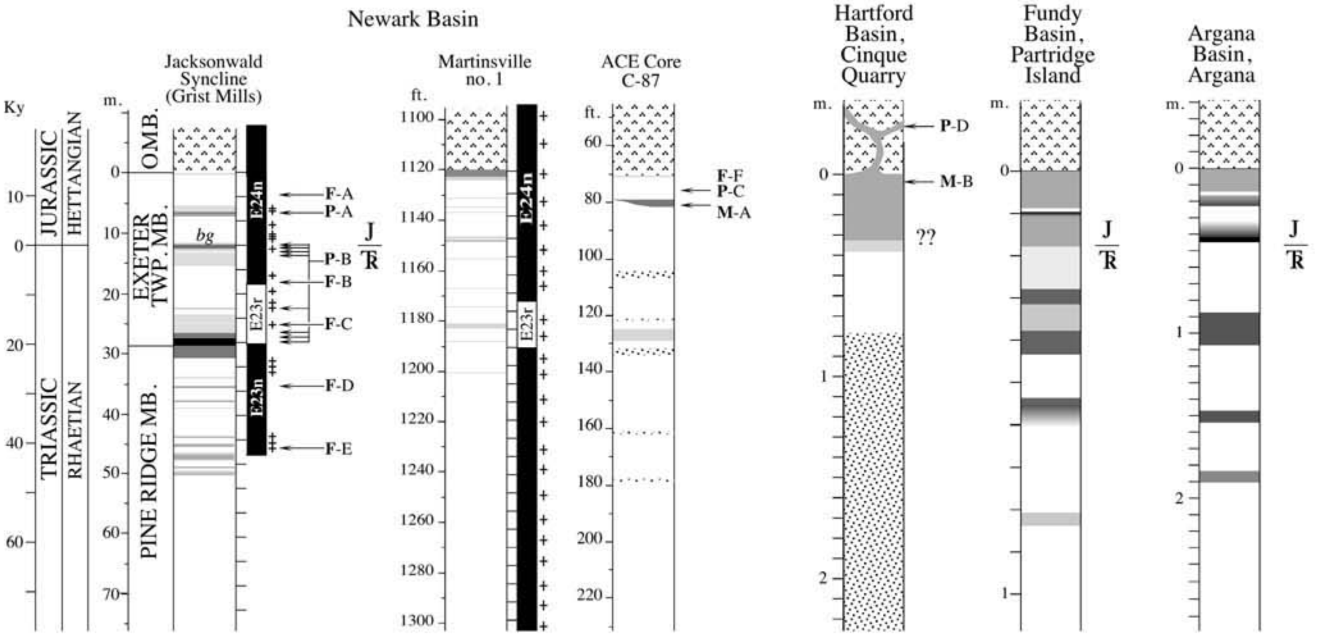

Figure 7. Detailed stratigraphy of the Newark, Hartford, Fundy, and Argana basin boundaries. In all cases the Triassic-Jurassic boundary is based on palynomorph transitions. Newark section from Olsen et al., 2001b. Abbreviations are: F-A, highest footprint bearing level in the Jacksonwald syncline with small grallatorids only (projected from Exeter Village sections); F-B, footprint assemblage with Batrachopus cf. B. deweyii, Rhynchosauroides, and small grallatorids (projected from Exeter Village sections); F-C, highest footprint level with new taxon B and Brachychirotherium and cf. Apatopus, also with Rhynchosauroides and small grallatorids; F-D, Brachychirotherium, new taxon B, Batrachopus, and small grallatorids; F-E, level with abundant Brachychirotherium new taxon B, Batrachopus, and small to medium sized grallatorids; F-F, Clifton-Patterson area quarries and exposures with small to large grallatorids, including the lowest occurrence of Eubrontes giganteus, also with Batrachopus deweyii, and Rhynchosauroides; P-A, lowest definitive Jurassic-type palynomorph level; P-B, palynomorph assemblages of Triassic aspect (lower) or dominated by spores (upper); P-C, palynomorph assemblage with Corollina only (poorly preserved); P-D, palynomorph assemblage with Corollina only in matrix between basalt pillow (Robbins in Heilman, 1987); M-A, macrofossil plant assemblage dominated by Brachyphyllum and Clathropteris; M-B, macrofossil plant assemblage dominated by Brachyphyllum (Heilman, 1987). bg, indicates position of the "blue-gray sandstone".

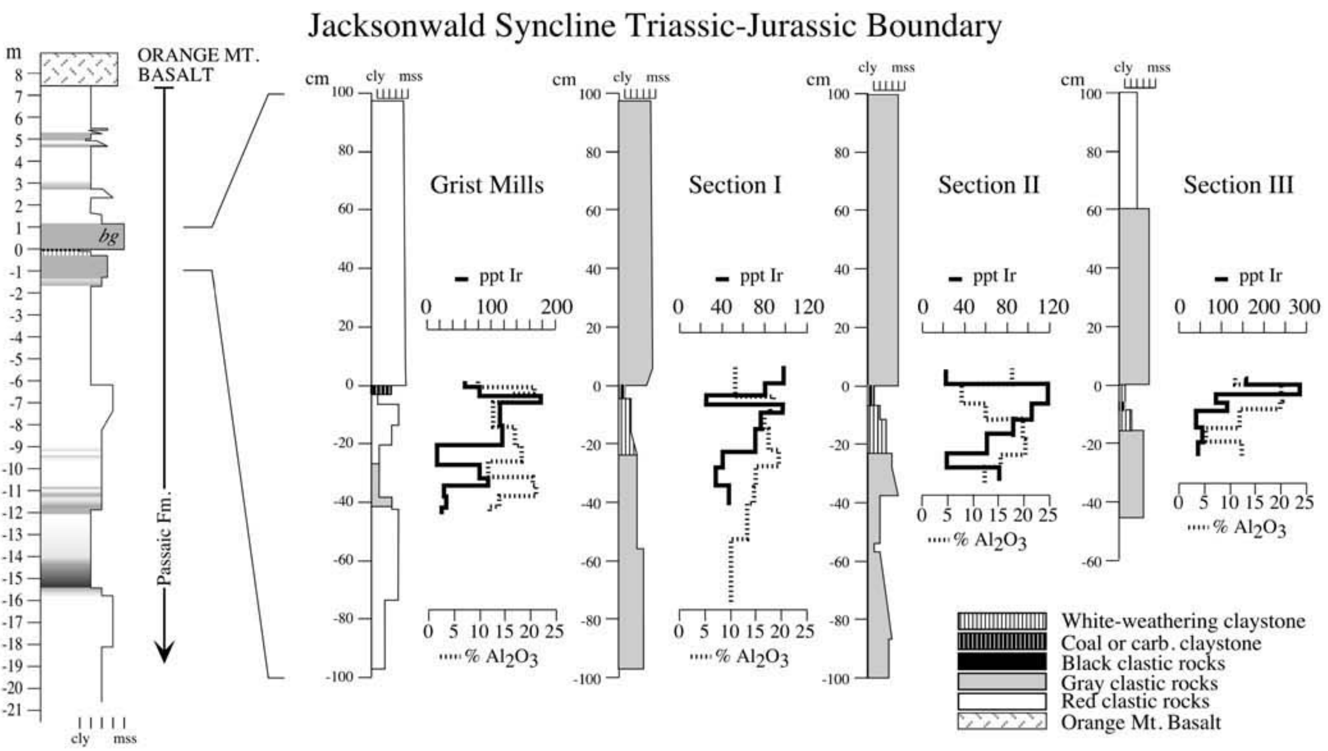

Figure 8. Iridium and aluminum concentrations at the four sections along strike in the Jacksonwald syncline (Fig. 3). Section on left is the average section (Fig. 9), and bg, indicates position of the "blue-gray sandstone".

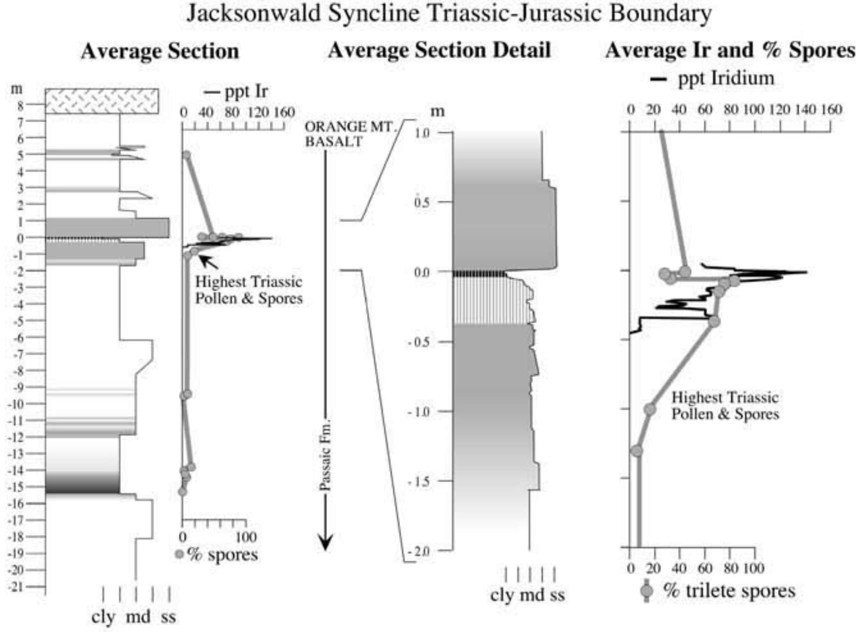

Figure 9. Summary of average Ir and pollen and spore data from the Jacksonwald syncline. Color, grainsize, and Ir are average values from the four sections shown in figure 8; pollen and spore data are from Fowell (1994). Key to lithologies as in figure 8. Color, Ir, and spore and pollen data, based on the interval ("channel") samples, were interpolated to a common (1 cm) sampling interval and averaged across the same depth levels for all localities. Spore and pollen data from Fowell (1994). Common datum for averaging is the base of the "blue-gray sandstone" at 0 m.

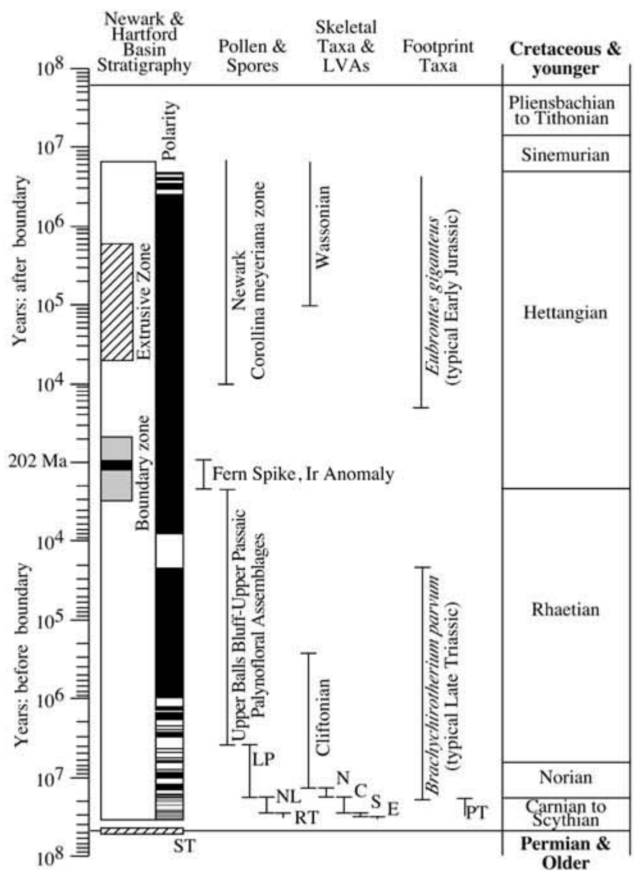

Figure 10. Summary of major physical and biotic events around the Triassic-Jurassic boundary plotted on a logarithmic scale. Data primarily from the Newark basin; time scale based on figure 4. LVAs are Land Vertebrate Ages of Huber et al. (1996) and Lucas and Huber (2001): N, Neshanician; C, Conewagian; S, Sanfordian; and E, Economian. Pollen and spore zones are from Cornet (1977) and Cornet and Olsen (1985): LP, Lower Passaic Heidlersburg; NL, New Oxford, Lockatong; RT, Richmond Taylorsville. Footprint distribution is from Olsen et al (2001b); PT, range of the Pekin-type footprint assemblages (Olsen and Huber, 1996). Note that the extrusive zone consists of lava flow formations interbedded with fossiliferous and cyclical sedimentary strata, with the latter interpreted as representing nearly all of the time shown. ST indicates the position of the Permo-Triassic Siberian Traps.