The Newark Basin, The Central Atlantic Magmatic Province, and the Triassic-Jurassic Boundary

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INTRODUCTION

This field guide is intended as an introduction to the rich record of Atlantic margin rifting preserved in the Triassic-Jurassic Newark basin (Figures 1,2). The superb temporal control and resolution provided by the Newark basin astronomically-calibrated paleomagnetic polarity timescale (NBTS) produced as a result of NSF-funded Newark Basin Coring Project (NBCP; 1989-1994) makes the Newark Supergroup, particularly the Newark basin, one of the best venues for examining rifting and related tectonic and biotic processes (Kent et al., 1995; Kent and Olsen, 1999; Olsen et al., 1996a, b, Olsen and Kent, 1999) (Figure 3). We will visit six stops (Figures 2,4) that illustrate this region's range of sedimentary and igneous environments, focusing on their significance to the understanding of global events in the early Mesozoic, especially in relation to the emplacement of the giant Central Atlantic Magmatic Province (CAMP) flood basalts and associated intrusives, the opening of the central Atlantic Ocean, and the Triassic-Jurassic mass extinction (Figure 5).[1]

Summary of Newark Basin Tectonic History

The latest Paleozoic and Mesozoic tectonic history of eastern North America, namely the type "Atlantic passive margin", including the Newark basin, is much more complicated and interesting than previously thought (e.g. Malinconico, 2003; Olsen, 1997; Schlische, et al., 2002; Wintsch, 1997; Withjack et al., 1995, 1998). It has long been realized that regional extension followed the late Paleozoic compressive and oblique assembly of Pangea relatively quickly. However, the discovery of Late Permian post-orogenic strata in Eastern North America and Morocco (Olsen et al., 2000) requires the initiation of deposition, probably in extensional basins, perhaps due to orogenic collapse, very soon after the Late Carboniferous-Early Permian docking of Africa with North America. A depositional hiatus of perhaps 15 million years between these Permian deposits and the oldest Triassic strata suggests a temporary cessation of plate divergence. Extension began in earnest in the Middle Triassic (~235 Ma), marked by the formation of many relatively small basins that filled with lacustrine and fluvial strata. At about 226 Ma (NBTS), a larger pulse of extension coalesced and enlarged many of these smaller basins, resulting in deposition of lacustrine and fluvial sediments in much larger basins (Stops 1 - 3). This was followed by the last known major pulse of extension at about 200 Ma, which resulted in the emplacement of the intrusions and extrusions of the CAMP, and the deposition of interbedded and younger lacustrine and fluvial sediments (Stops 4 – 6). All of these NW-SE-directed extensional pulses reactivated suitably-oriented Paleozoic compressional faults along which the rift basins formed, following the pattern demonstrated by Ratcliffe (1971) and Ratcliffe et al. (1986).

One of the largest unresolved geodynamic questions involving these series of rifts, including the Newark, is the relationship between the CAMP and the initiation of

[1] Parts of this guidbook are drawn from two other guidebooks, previously written: Olsen, P.E. and Rainforth, E.C., 2003, The "Age of Dinosaurs" in the Newark basin, with special reference to the lower Hudson Valley. in Gates, A.E., (ed.) and Olsen, P.E., (organizer), Geology of the Lower Hudson Valley, 2001 New York State Geological Association Field Trip Guide Book, New York State Geological Association, New York State Museum, Albany, v. 73 (2001), p. 59-176; and Olsen, P. E., 2003, Field Guide for Non-marine Boundary Events in the Newark Basin (New Jersey, Pennsylvania, and Connecticut), Eastern United States and their Litho-, Chrono- and Biostratigraphic Context. Guidebooks for Field Workshops of IGCP.

seafloor spreading. During the early development of modern plate tectonic theory, it was widely assumed that the igneous activity now known as CAMP was synchronous with the initial formation of oceanic crust in the central Atlantic (e.g., Phillips and Forsyth, 1972; Pitman and Talwani, 1972; Dallmeyer, 1975) with vestiges of this point of view persisting until quite recently (e.g. Holbrook and Keleman, 1993). However, two independent sets of analyses have led to the current conventional view that the CAMP and related continental rifting are essentially decoupled from initial seafloor spreading. First, extrapolation of marine magnetic anomaly and fracture zone data has suggested that drift began between 190 and 170 Ma (Klitgord and Schouten, 1986; Benson, 2003), and was intimately related to the emplacement of seaward dipping reflectors (sdr's), that are convincingly argued to be a massive volcanic wedge (Holbrook and Keleman, 1993; Talwani et al., 1995; Oh et al, 1995). Second, more precise age assessment of the CAMP by a variety of techniques, has shown that its age is between 202 and 195 Ma (Sutter, 1988; Olsen et al., 1996b; 2003; Marzoli et al., 1999; Hames et al., 2000) and therefore as much as 30 m.y. older than the onset of drift.

In contrast, Withjack et al. (1988) and Schlische et al., (2003) argue that the initiation of drift was fundamentally diachronous in the Central Atlantic region and that the CAMP acts as a temporal benchmark for that diachroneity. In their scenario, continental rifting ceased before or at the Triassic-Jurassic boundary in the southeastern US coeval with the emplacement of the earliest Jurassic CAMP as dikes, flows, and the southern segment of the sdr's. The CAMP emplacement occurred during regional compression coaxial with the previous extension direction (i.e., tectonic inversion) and was post-rift. In the northeastern US, such as in the Newark basin (Stops 1-6), Maritime Canada and Morocco, however, continental rifting actually accelerated during earliest Jurassic CAMP emplacement, and tectonic inversion was substantially later in the late Early to Middle Jurassic coeval with the initiation of drift and emplacement of younger sdr's in that region (c.f., Oh et al., 1995). This was the second of similar sequences of events through the Mesozoic and Cenozoic that progressed from south to north along the mid- to North Atlantic margin along crustal segments. These segments are delineated by major strike-slip plate boundaries inherited from the Paleozoic assembly of Pangea. The segment containing the Newark, Hartford, Fundy and most of the Moroccan rift basins is bound by the Minas Fault – Gibraltar Fault zone on the north and the Long Island Platform boundary – South Atlas fault zone on the south (Figure 1).

These three different schemes for the evolution of the onset of seafloor spreading have rather different implications for underlying geodynamic mechanisms, with "passive", slab-pull driven plate motion being compatible with the early and present conventional views of Central Atlantic opening, while "active" asthenospheric upwelling is more compatible with the model of Withjack et al. (1988) and Schlische et al., (2003).

In the Newark basin, it appears that after more than 30 m.y. years of extension in response to divergent plate motion, compression set in coaxial with the original extension direction. The evidence is as follows: 1, dramatically higher thermal maturities (Malinconico, 2002) and reset zircon fission-track ages (Hoek et al., 1998; Steckler et al., 1993) along the eastern margin of the presently-delineated Newark basin, suggest significant post-rift uplift and much more erosion of the eastern relative to western basin margin (Stops 1-3); 2, much if not most of the westward tilt of the Newark basin strata occurred post-depositionally, based on a pervasive mid-Jurassic paleomagnetic overprint, which itself also suggests a major mid-Jurassic fluid-flow event (Kent et al., 1995; Witte and Kent, 1991); 3, pervasive small-scale bedding-plane faults that show consistent reverse (in present coordinate space) motion based on offset of bedding-normal joints, small scale folds and thrust faults, and slickenline orientations that appear, on the basis of our casual observations, to be coaxial with the inferred older extension direction, thus requiring post-depositional NW tilting and NW-SE shortening in an approximately horizontal plane (Stop 2). These observations and interpretations are consistent with the basin inversion geometry described by Withjack et al. (1998). This geometry is based on physical models that produce a major up-arching of the basin during compression, but with little reverse motion on the boundary faults and only subtle compressional deformation of the basin fill. Erosion of several km of basin section and surrounding basement rocks took place during this late Early to Middle Jurassic tectonic inversion event.

By the Early Cretaceous the present erosional level had been reached in the Newark basin, and marine and marginal-marine Cretaceous and Cenozoic coastal plain deposits onlapped the rift deposits, although according to Wintsch (1997) uplift continued well into the Cretaceous in the Hartford basin. Northwest-southeast compression slowed sometime in the Cretaceous or Cenozoic and switched to the NE-SW horizontal compression that persists to the present day (Goldberg et al., 2002; Zoback, 1992; Zoback and Zoback, 1989) perhaps reflecting the most recent phase of tectonic inversion related to the far field effects of seafloor spreading in the North Atlantic.

The Triassic-Jurassic Boundary Event

An additional unresolved provocative problem is the relationship between the Triassic-Jurassic mass extinctions (Figure 3) and the CAMP (and by extension seafloor spreading). In eastern North America the Triassic-Jurassic extinction level lies below the oldest CAMP basalts, and a thin intervening interval of earliest Jurassic age strata with a stereotyped fauna and flora (Stop 4). A similar fauna and flora characterize the strata interbedded with and overlying the succeeding CAMP basalt flow formations (Stop 6). Based on Milankovitch cyclostratigraphy that is consistent with available radiometric dates, this pre-CAMP Jurassic interval lasted for 20 to 40 ky (Olsen et al., 1996a). In the Newark basin, the extinction level is itself characterized by a fern spore spike, modest Ir anomaly (Olsen et al., 2003), and strong negative $\delta^{13}C$ excursion (Whiteside et al., 2003). While superposition dictates that the CAMP must follow the extinctions in eastern North America and the evidence from the boundary itself is suggestive of an impact origin, many authors have argued that the CAMP probably erupted earlier elsewhere and actually caused the extinctions via CO_2-induced climate change (Courtillot, 1994, 1999; McElwain, 1999; Beerling and Berner, 2002; Hesselbo et al., 2002;).

A possible causal link between impact, CAMP, and the Triassic-Jurassic extinctions is difficult to ignore, given a similar (although less precisely timed) coincidence between the Deccan Traps and the K-T boundary and the Siberian Traps and the Permo-Triassic boundary (Rampino and Caldeira, 1993; Becker et al., 2004) (Figure 6). The three largest Phanerozoic mass-extinctions (P-Tr, Tr-J, and K-T) are thus closely penecontemporaneous with the three largest Phanerozoic flood basalt provinces (Siberian, CAMP, and Deccan), and there is at least some evidence of an asteroid or comet impact. An early suggestion by Dietz (1986) was that an impact generated a hot spot at the “Bahama Nexus” causing the entire CAMP event and initiating drift. Boslough et al. (1996) proposed a mechanism linking bolides with flood basalts, but the energetics have yet to be reconciled with the observations and the models (Melosh, 2000; Ivanov and Melosh, 2003). Although highly speculative, it seems plausible that a massive impact might be able to concentrate the effusive rate of a distant flood basalt province, but this topic has yet to be explored quantitatively at the appropriate scale.

FIELD TRIP STOPS

Stop 1: Palisades Interstate Park (1 hr)

Major features:

1) Early Jurassic Palisade sill of the CAMP
2) Hinge margin of basin
3) Late Triassic Stockton and Lockatong Formations

Context

This stop is along the hinge margin of the Newark basin. Looking east, the contact between the rift basin fill and metamorphic basement is approximately in the middle of the river with about 230 m of intervening sedimentary strata dipping gently to the west. The sedimentary strata exposed here consist of the Nursery and Princeton members of the Lockatong Formation of Late Triassic age (219-220 Ma; NBTS) and upper Stockton Formation, the latter probably correlative to the lowest Lockatong deeper into the basin. In the central part of the basin, correlative strata lay 1800 m above basement, with the thinning being due to a combination of hanging wall onlap and stratal thinning towards the apex of the half-graben sedimentary wedge. The main point of this stop, however, is the Palisade sill, which intrudes the lower Lockatong Formation in this area and tends to be highly concordant to bedding,

The Palisade sill is one of the most famous intrusive bodies in the world. It has been the subject of numerous papers, some classic (e.g., Darton, 1889, 1890; Lewis, 1908; Walker, 1940; Walker, 1969) and is a classic field stop for geological excursions (Puffer, 1987). Not appreciated until recently, however, is the immense total size of the body and its relationship to what is now called the Central Atlantic Magmatic Province (CAMP) (Marzoli, 1999). Its injection marks the initiation of flood basalt activity in the basin and probably in the North American plate.

The overall composition of the sill a light REE-enriched, high-Ti, quartz tholeiite ranging texturally from a fine-grained diabase to a coarse gabbro. It is compositionally very similar to Karoo basalts It was intruded in the earliest Jurassic at about 200 Ma [201.5±0.5 Ma (U-Pb: Dunning and Hodych, 1990); 201±0.6 Ma (^{40}Ar/^{39}Ar -Turin, 2000); 200.4±2.6 Ma, 200±1 Ma (^{40}Ar/^{39}Ar - Baksi, 2003) and fed the initial basalt flows (Orange Mountain Basalt) of the Newark basin of indistinguishable ^{40}Ar/^{39}Ar age (201±2.1 Ma).

The sill is an average of about 300 m thick, and at outcrops in this region there is a 3 to 4 m thick zone rich in olivine about 12 - 16 m above the chilled contact with the underlying metamorphosed Lockatong Formation of Late Triassic age (219 Ma; NBTS). The olivine zone was interpreted by (F. Walker, 1940) as a part of what became known as the Palisades olivine-clinopyroxene fractionation series resulting from gravitation crystal settling. K.R.Walker (1969), however, showed the olivine is too fine-grained to have settled far through the magma and it is also too Fe-rich. Instead, he suggested that the olivine zone represents a second injection of olivine basalt into the previously intruded, partially crystallized mush, followed by gravitational settling of the largest crystals. Shirley (1987) proposed a third injection into the layered sheet in this region. The upper part of the sill, below the upper chill margin tends to be granophyric, although there are thin dikes of granopyhre near the base of the sill.

The sill itself, however, is only part of a series of probably once-continuous sills and plutons that extended over almost the entire Newark basin, making this component of the CAMP one of the most extensive sills in the world. Other parts of this sill-pluton-complex have bronzite layers instead of olivine and it seems clear that there was lateral transport of fractionating magma in multiple directions.

An important aspect of the sill emplacement is that while the basin fill had to expand by the thickness of the sill (ca. 300 m) in the time the sill would take to cool (10s of ky). Some of this may have been compensated for by water loss in the sediments during metamorphism, but one might expect inflation and elevation of the basin surface. Yet, based on Milankovitch cyclostratigraphy from the NBCP and other cores the accumulation rate and maximum water depth was considerably higher immediately after the extrusion of the earliest Jurassic Orange Mountain basalt (Schlische and Olsen, 1990; Olsen et al., 1996a), suggesting an increase in accommodation space. On the other hand, mass was transferred from some magma chamber at depth to the sill and to the surface, which could potentially lower the surface, but the wave length of the subsiding area would presumably be much greater than the width of the basin. Consequently, we suppose that the emplacement of the sill and extrusion of the lavas, the deepening of the lakes and the increase in accumulation rate were all a consequence of a very large increase in extension that probably began slightly before the igneous event itself.

<u>Stop 1a: Palisade sill:</u> 40°51.460'N, 73°57.534'W.

This old quarry, at the traffic circle at the ramp leading to Ross Dock, exposes the lower half of the Palisade sill (Figure 7). As noted by Walker (1969), the olivine zone of the sill produces an obvious bench along the escarpment to the immediate south and elsewhere in the area, essentially paralleling the lower contact of the sill. In the cliff face, the olivine zone is marked by a zone of deflected columns. Flow banding is present within the olivine zone (Naslund, 1998).

Proceed south along Henry Hudson Drive.

<u>Stop 1b: Concordant contact between Palisade sill and Lockatong Formation:</u>
40°51.253'N, 073°57.587'W.

This comprises what is probably the most spectacular exposure anywhere in the basin of the Palisade sill with the underlying sedimentary rock (Figure 8). The contact is exposed for more than 50 m along strike, and reveals the intimate structure of the Lockatong-sill contact.

Note that the contact is extremely sharp and that there is virtually no evidence of assimilation of sediment into the sill. Here and there the sediment-sill contact jumps a few tens of centimeters up or down, but on the whole it is remarkably concordant.

The mudstones are metamorphosed here to a black to dark gray hornfels comprised of albite and biotite with minor analcime, diopside, diopside, and calcite and green hornfels made up of diopside, grossularite, chlorite, and calcite, with minor prehnite, amphibole, feldspar, and biotite (Van Houten, 1969; Puffer, 1987).

Proceed south along Henry Hudson Drive.

Stop 1c: Discordant contact between Palisade sill and Lockatong Formation: 40°51.063N, 073°5.672'W.

This exposure shows the contact between the Palisade sill and Lockatong Formation jumping first up several meters and then back down (Figure 9). The more mud-rich intervals appear to behave as if brittle, with essentially no assimilation into the sill, while the tan arkose appears to have flowed at the sill contact, with considerable chaotic mixing into the sill.

Proceed south along Henry Hudson Drive.

Stop 1d (optional): Outcrops of Stockton Formation along Hudson River shore: 40°50.886'N, 073°57.727'W to 40°50.802, 073°812'W.

From just south of here to beneath the George Washington Bridge there are scattered outcrops of variegated strata of the Stockton Formation; their facies are characteristic of the units immediately below the Lockatong in this region. Tan, gray, and purple trough cross-bedded arkose and pebbly arkose grade upwards, through a series of irregular beds, into bright purplish-red massive mudstone, at meter-scale repetitions. A large disarticulated postcranial skeleton of the archetypal Late Triassic semi-aquatic reptile, a phytosaur, was recovered on private land just south of where the path comes down to the water's edge (Figure 10). This specimen (AMNH 4991) is currently on exhibit in the Hall of Vertebrate Origins at the American Museum of Natural History in New York. The crocodile-like phytosaurs are perhaps the most common large tetrapod fossil in Late Triassic age strata in North America and they were abundant to the end of the Late Triassic, where they went extinct.

Walk back up hill to Henry Hudson Drive and proceed west toward intersection with Main Street (Fort Lee) entrance to the park.

Stop 1e: Weathered olivine zone: 40°50.798'N, 073°57.882'W.

Here a weathering profile of the olivine zone can be clearly seen. The olivine zone is weathered to a very crumbly diabase, that according to Naslund (1998) still has many fresh-looking olivine crystals.

Return to bus.

Stop 2: Granton Quarry (1 hr) 40° 48.431'N, 074 01.071'W (1 hr).

Major features:

1) Exposures of Late Triassic Lockatong Formation
2) Hinge margin of basin
3) Well-developed Milankovitch stratigraphy
4) Abundant and characteristic lacustrine fossils
5) Granton CAMP sill
6) Tectonic inversion and deep erosion

This locality is on the upper surface of the Palisade sill (Stop 1). Profoundly cyclical lacustrine strata of the Lockatong Formation are exposed at the abandoned Granton

Quarry on the hinge margin of the basin. Remnants of the old Granton Quarry are preserved between the new Lowes Home Building Center on the south and Tonnelle Plaza (Hartz Mountain Industries) on the north. Granton Quarry was actively quarried for road metal, fill and rip rap during the 1950s and 1960s and was abandoned by 1970, thereafter it was slowly consumed by commercial developments and warehouses. Nonetheless, excellent exposures remain of most of the Ewing Creek Member of the Lockatong (219 Ma, NBCP).

This site has produced, and continues to produce, extraordinarily abundant fossils, especially vertebrates, and it is certainly one of the richest sites in North America for the Triassic (Figure 11). This is also the best locality on this trip to see the details of Lockatong-type Milankovitch lake level cycles forced by the ~20 ky cycle of climatic precession (Figures 12, 13). These precession-forced cycles are called Van Houten cycles after their discoverer (Van Houten, 1962, 1964; Olsen, 1986; Olsen and Kent, 1996) and they are modulated in their expression by several orders of other cycles reflecting orbital eccentricity cycles (Figure 12). Eleven Van Houten cycles with a thin-bedded to laminated deep water portion are exposed on the sill-capped hill: seven are exposed on the south-facing exposure, three additional cycles are exposed on the east-facing exposure; and all 11 cycles are exposed on the north-facing exposure, which is where we will examine them. This section has been described in several papers including Van Houten (1969), Olsen (1980), Olsen et al. (1989), and Colbert and Olsen (2001).

The specific pattern of Van Houten cycles, distinctive marker beds, and fossil content allow these cycles to be tied to the cyclical sections below the Palisade sill and correlated to the NBCP cores (Figure 14). It is in the NBCP cores that the Milankovitch cyclicity is best displayed in what is still arguably the longest continental record (22 m.y.) of the full spectrum of precession related Milankovitch forcing recovered anywhere of any age to date. A representative power spectrum that includes the stratigraphic equivalent of this stop is shown in Figure 15.

With the sections from both sides of the Palisade sill combined and correlated with the NBCP cores, it is now possible to look at trends in lithology and biota at scales from a few thousand years (within one Van Houten cycle) to over 1 m.y. (Figure 14). The most obvious difference from the NBCP cores is that the Van Houten here cycles are thinner. In this area the Van Houten cycles are on average 1.5 m thick, but to the southwest towards the center of the basin, they average 4 m thick in the Princeton no. 1 core and about 5.1 m thick in the Nursery core. The correlative section in outcrop along the Delaware River, in the geographic center of the Newark basin, has Van Houten cycles that average 5.9 m thick. The thinning to the northeast reflects deposition closer to the hinge margin of the basin. An additional change is that there tends to be some omission of Van Houten cycles in eccentricity minima along the hinge margin, which almost certainly reflects a cessation of deposition during drier times along the basin edge.

Although each cycle has its unique properties, there are prominent general paleontological patterns repeated in most Van Houten cycles, which are well shown in cycles G7 and G3 (Olsen et al., 1989). The most obvious and least surprising pattern is seen in the correlation between the degree of fish preservation and the degree of lamination of the sediments. Microlaminated beds tend to preserve beautifully articulated fish, laminated mudstones produce disarticulated but still associated fish, and mudcracked mudstones have only dissociated scales and skull bones. This correlation almost certainly reflects the often-quoted dependence of fish preservation on a lack of oxygen, bioturbation, macro-scavenging, and physical disturbance.

Inversely correlated with this fish-preservation trend is one that at first appears very peculiar: a tendency for lower fish diversity in the beds with the best fish preservation, and vice versa. This trend is all the more surprising because many more fish (individuals) are identifiable from the beds producing the best-preserved fish. The explanation seems to be that the highest diversity of lake environments tends to be near the shores, whereas the deeper-water zones tend toward low diversity (this is especially true for lakes with anoxic bottom waters, because they lack benthic forms). Thus, the cycle in fish diversity is a consequence of proximity to the shore that varies as a function of depth, which in turn controls the degree of lamination and absence of bioturbation.

The taphonomic pattern seen in the microlaminated deep-water phase of Van Houten cycles fits a chemically-stratified lake model (Bradley, 1929, 1963; Ludlam, 1969), in which bioturbation is perennially absent from the deeper parts of the lake bottom because the bottom waters lack oxygen, which is required by almost all macroscopic benthic organisms. Chemical stratification (meromixis) can arise by a number of mechanisms, but the main physical principle involved is the exclusion of turbulence from the lower reaches of a water column. This tremendously decreases the rate at which oxygen diffuses down from the surface waters, and retards the upward movement of other substances. The main source of water turbulence is wind-driven wave mixing. This turbulence usually extends down about one-half the wavelength of surface wind waves, which depends on the fetch of the lake, wind speed, and wind duration. If the lake is deeper than the depth of the turbulent zone, the lake becomes stratified with a lower non-turbulent zone and an upper, turbulently-mixed zone. The thickness of the upper mixed zone is also dependent on density differences between the upper waters (epilimnion) and lower waters (hypolimnion), which can be set up by salinity differences (saline meromixis) or by temperature differences, as seen in many temperate lakes. In the absence of saline or temperature stratification, chemical stratification can still arise in a deep lake with relatively high levels of organic productivity. Because oxygen is supplied slowly by diffusion, consumption by bacteria of abundant organic matter sinking into the hypolimnion plus oxidation of bacterial by-products eliminates oxygen from the hypolimnion. Lakes Tanganyika and Malawi in East Africa are excellent examples of very deep lakes in which there is very little temperature or density difference between the epilimnion and hypolimnion, but chemical stratification still occurs with the exclusion of oxygen below 200 m. Such a pattern is common in deep tropical lakes. The preservation of microlaminae and fossils in Van Houten cycles may have been a function of great water depth relative to a small surface area of the lake, which in the case of Lockatong lakes was none the less huge (in excess of 10,000 km^2); the depth, based on the area of the lake that must have been below the turbulent zone, was a minimum of 80 m during the deposition of the microlaminated beds (Olsen, 1990).

Thus, Van Houten cycles with a microlaminated deep-water phase, such as G3 and G7 at Granton Quarry reflect the alternation of shallow, ephemeral lakes or subaerial flats with deep perennial lakes with an anoxic hypolimnion set up by turbulent stratification under conditions of relatively high primary productivity and low organic consumption (e.g. low ecosystem efficiency). The low organic content of the transgressive and low-stand portions of Van Houten cycles in general probably reflects higher ecosystem efficiency caused by shallow water depths, rather than lower total organic productivity.

At this stop, cycles G3 and G7 have produced representatives of all the known skeletal remains of Lockatong vertebrates except the holostean *Semionotus* (Figure 11A).

The basal deep-water portions of both of these cycles have extremely high densities of fossil fish, especially the coelacanth *Osteopleurus newarki* Schaeffer (1952). Small reptiles are also surprisingly abundant, and many important fish and unique reptile skeletons have been discovered here by dedicated amateurs and donated to various museums through the years (Colbert, 1965, 1966; Colbert and Olsen, 2001; Olsen et al., 1989; Schaeffer, 1952; Schaeffer and Mangus, 1970. Without a doubt the three most spectacular skeletons of small reptiles found in the Lockatong come from the Granton Quarry. These include the type specimen of the bizarre "deep-tailed swimmer", *Hypuronector limnaios* (Colbert and Olsen, 2001), the peculiarly-abundant tanystropheid *Tanytrachelos ahynis* (Olsen, 1979), and the gliding lepidosauromorph *Icarosaurus seifkeri* (Colbert, 1966) (Figure 11B). Larger remains occur as well, of which the most spectacular is the skull of a juvenile rutiodontine phytosaur, but isolated phytosaur bones and teeth are fairly common and isolated vertebrae of a metoposaur amphibian also been found.

At the south-facing exposures, cycles G1 and G2 are injected by diabase of the 20 m thick Granton sill (Van Houten, 1969), another component of the CAMP, which has protected the Lockatong Formation from erosion in this area. Notice the absence of prominent folding at the diabase-sediment contact. Because this sill is thin, and the Palisade sill fairly remote, much of the sedimentary rock is not as metamorphosed as at previous stops; some cycles still have considerable organic matter.

According to Van Houten (1969), these Lockatong hornfels include calc-silicate varieties in the middle carbonate-rich part, and extensively feldspathized and recrystalized diopside-rich arkose in the upper part. Some beds of arkose show well-developed cross-bedding. Because of the buff arkose at the top of nearly every cycle, these are the most visually graphic of the detrital cycles seen on this field trip; here the many correlated changes occurring though individual cycles can be easily seen (Figure 13).

One or two bedding plane thrust faults, always thrusting to the east, are present in deeper-water phase of nearly every cycle at Granton Quarry. Slickensides are usually present and indicate that movement occurred parallel to dip. All the joint sets are cut by these thrusts, their displacement indicating that each fault has a net slip of 0.5 to 1.5 cm. This type of minor thrust fault is evident in virtually all Newark Supergroup lacustrine cycles and can be seen at every stop of this trip. The fact that all of these faults are thrusts requires post-depositional northwest-southeast shortening, a steepening of dip, and a $\sigma 3$ that would have been vertical. This is completely incompatible with the extension that produced the basins; thus we take these faults as evidence for structural inversion – shortening coaxial with the original tilting.

Return to bus.

Stop 3: Kilmer Member of Passaic Formation at Lyndhurst, NJ. 40°48.434'N, 074°06.506'W. Lunch (1 hr)

Major features:

1) Exposures of Late Triassic middle Passaic Formation
2) Widening basin
3) Marginal lacustrine facies
4) Late Norian fauna

We have moved considerably up section from Stops 1 and 2 as well as deeper towards the depositional center of the basin, although we are still in a marginal facies. These exposures reveal most of the Kilmer Member of the Passaic Formation (210 Ma, NBTS) (Figure 2B) exposed on the east side of a prominent ridge marking the western border of the New Jersey Meadowlands. This ridge is characterized by a heterogeneous assemblage of red mudstones and sandstones. However, there are a few purple and gray units present, and the eastern and stratigraphically lowest of these mark the base of the Kilmer Member (Figure 16).

The Passaic Formation marks a stage in the tectonic evolution of the Newark rifts characterized by extremely widespread units with a particularly well-developed "layer-cake" style stratigraphy. Based on the very slow rates of thickening towards the faulted margins of the basin, extension rates were slowing, the basin was filling towards its outlet, and the basin floors were extraordinarily wide and flat. The red Passaic and other central Pangean equivalents are the strata that most people think of when they think Triassic. At yet broader scale, the Passaic marks an interval that in basins deeper within the arid belt and closer to the rifting axis began to receive considerable amounts of brine of marine origin with the consequent development of thick evaporite sequences (e.g. Osprey Salt of the Canadian Maritime margin).

The exposures at Stop 3 reveal the uppermost few meters of member T-U and the lower half of the Kilmer Member (Figure 2). In the central Newark basin, the basal Kilmer Member includes a prominent Van Houten cycle with a well-developed black division 2. In the region around New Brunswick (NJ), this black shale and the underlying division 1 of this cycle are often rich in copper minerals, particularly calcopyrite. At these outcrops in the northeastern Newark basin, the same Van Houten cycle apparently lacks black shale, instead having a purple shale with associated tan or white sandstones. The unit is still copper-mineralized, at least locally, and where intruded by thin diabase sills 2 km to the south-southeast in North Arlington (NJ), it was commercially exploited in what is supposed to be the oldest copper mine in North America – the Schuyler mine (Lewis, 1907). The exposures at this stop may be the prospect mentioned by Woodward (1944) on the Kingsland estate, inspired by the Schuyler mine, but never worked extensively. An exploratory shaft at least was opened, and the now-cemented entrance is still visible. Tan and white sandstones associated with purple and gray mudstone are exposed, and mineralized with the same minerals as at the Schuyler mine, including chalcocite (black copper sulfide), chrysocolla (bluish-green copper silicate), malachite (green copper carbonate), and azurite (blue copper carbonate).

The overall section at this stop consists of lower red massive mudstones of member T-U, followed by the tan and white sandstones surrounding a purple well-bedded mudstone of the basal part of the Kilmer Member. This is succeeded by massive red mudstones, a well-bedded interval, and then red mudstones and fine sandstones with gypsum nodules. The overall stratigraphy is very similar to the expression of member T-U and the Kilmer member in the NBCP cores.

A large collection of very well-preserved reptile footprints was made near here by Lawrence Blackbeer in the late 1960s (pers. comm., 1985; Olsen and Baird, 1986) (Figure 17). Although the exact location was not recorded in detail, the lithology of the footprint slabs is consistent with the local expression of the Kilmer Member. This was confirmed by the discovery in 2002 of several well-preserved tracks in appropriate lithology from between the two prominent copper-bearing sandstones at this spot. This late Norian age assemblage is distinguished in the Newark basin by the presence of relatively large grallatorid (theropod dinosaur) footprints, up to the size of *Anchisauripus tuberosus*; this is the oldest level in the basin with such tracks. Relatively large examples of the ornithischian dinosaur track, *Atreipus milfo*rdensis, are present, along with a new dinosaurian ichnogenus ("*Coelurosaurichnus*" sp. of Olsen and Flynn, 1989), as well as the non-dinosaurian *Brachychirotherium parvum*, and *Rhynchosauroides* sp. This track assemblage records the beginning of the rise to ecological dominance of the dinosaurs, which in lower horizons are conspicuous by their relative rarity and small size.

Return to Polito Avenue and turn left, heading north.

Stop 4: Triassic-Jurassic boundary and Overlying Initial CAMP Basalt. Tilcon Quarry, Clifton, NJ. 40°52.470'N, 074°11.382'W. (1.5 hr).

Major features:

1) Exposures of uppermost Passaic Formation and overlying Orange Mountain Basalt of the CAMP
2) Possible minor unconformity caused by increasing extension
3) Triassic-Jurassic boundary
4) Earliest Jurassic recovery fauna and flora

The Tilcon Quarry (formerly H.R. Hamilton Quarry) in Clifton, NJ is winding down its activity and is scheduled for development as a residential community (Figure 2). However, for the last 20 years it has been a preeminent locality for the amateur collection of thousands of dinosaur and other footprints. Within the last few years, it has become apparent this site exposes a particularly interesting facies of the Triassic-Jurassic boundary and its tectonic milieu. The quarry exposes about 20 m of uppermost Passaic Formation and most of the 55 m thick lowest of three major flows of the 150 m thick Orange Mountain Basalt (Figure 18).

In this area, the uppermost Passaic Formation consists of two units of strongly contrasting facies: a lower interval of fluvial facies, and an upper unit of marginal lacustrine facies (Figure 19). The bulk of the upper Passaic in this region consists of upward fining cycles of relatively poorly sorted pale red conglomerate and pebbly sandstone with poorly defined trough cross bedding grading upward into massive red mudstones and sandstones (described by Parker et al., 1988). The units are intensely bioturbated, which is what obscures the bedding. This facies is overlain with a sharp contact by red and gray mudstones and sandstones that have much more distinct bedding and excellent preservation of small-scale sedimentary structures, and which are overtly more heterogeneous than underlying units. There are cross-laminated sandstones with channel morphologies, tilted thin beds that toe laterally into mudstones suggestive of small deltas, tabular beds of climbing ripple cross-lamination, and thin bedded mudstone beds suggestive of suspension deposits in standing water. Many sandstone beds are bound by clay drapes, and many surfaces are covered by ripples, desiccation cracks, and trace fossils, notably reptile footprints. The uppermost surface of this interval is covered by the Orange Mountain basalt and locally seems to have preserved some depositional relief, including small channels (Figure 18). Floral and vertebrate remains from the upper facies completely lack all Triassic forms indicating a Jurassic age.

The uppermost Passaic Formation contains the Triassic-Jurassic boundary as identified by an abrupt extinction level in pollen and spores (Fowell and Olsen, 1993) (Figure 20). In the southeastern Newark basin, where accumulation rates are highest, and the Milankovitch stratigraphy around the boundary most clearly expressed, the palynological extinction level is directly associated with a tetrapod extinction event, “fern spike”, modest Ir anomaly, and negative $\delta^{13}C$ anomaly (Olsen et al., 2002; Whiteside et al., 2003), observations we have used to support an impact origin for the extinctions.]

There is a thin zone of reversed polarity (E23r) just below the boundary (Kent et al., 1995; Kent and Olsen, 1999) (Figures 2 and 19). This same reversed zone was encountered in homotaxial, although muted, cyclical sequence in the Martinsville no. 1 NBCP core (Figure 19). In the outcrops in southeastern Pennsylvania and the Martinsville no. 1 core, there is essentially no facies change across the boundary, with the entire section being lacustrine.

The same set of facies seen at Stop 4 is exposed 1.5 km to the south on the grounds of Montclair State University, which was also well known as a major dinosaur footprint site. The coring transect of the Army Corps of Engineers (ACE) Passaic River Diversionary Tunnel Project passed through this area and a series of short cores (<200 m) were recovered through the uppermost Passaic and lower Orange Mountain Basalt (Fedosh and Smoot, 1988; Olsen et al, 1996a; Olsen et al., 2002) all of which show the same facies transition as seen at Stop 4 (Figure 19). Thus far, we have been unable to find this reversed interval E23r in our preliminary work on the ACE cores, although we stress that we have not yet sampled the quarry section.

The apparent absence of E23r coupled with the abrupt facies change seen here and in the ACE cores suggests that these outcrops close to the northeastern terminus of the Newark basin have a minor unconformity at the boundary. This unconformity would be a tectonostratigraphic sequence boundary that would pass into a correlative conformity deeper in the basin (c.f., Olsen, 1997). We interpret this tectonostratigraphic sequence boundary as being consistent with an increase in extension rate, and increasing tilting rate, the expression of which is the much higher accumulation rates and deeper lakes present in the overlying sedimentary units (Feltville, Towaco, and Boonton: Olsen et al., 1996a). The upper facies at this locality represents an onlap of marginal lacustrine Jurassic strata directly onto eroded Triassic strata.

The floral assemblage from the upper facies of Stop 4 include abundant remains of *Brachyphyllum*, a cheirolepidiaceous conifer and less common fragments of the dipteridaceous fern *Clathropteris meniscoides*, a large variety of casts of stems and rizoliths, and poorly preserved pollen mostly of the genus *Corollina* (Figure 19). This is a typical earliest Jurassic assemblage very similar to that in the overlying Feltville Formation.

However, the faunal remains are truly spectacular (Figure 21), consisting of very numerous and often well-preserved footprints. These earliest-known Jurassic track assemblages are of strikingly low diversity, despite collection and identification of thousands of specimens. Only the lizard-like *Rhynchosauroides* sp., the crocodiliomorph *Batrachopus deweyii*, and the theropod dinosaur tracks *“Grallator”* spp., *“Anchisauripus”* spp., and *Eubrontes giganteus* are known from these beds. Similar tracks have been found in the upper facies at other near-by localities such as Montclair State University. Despite the thousands of tracks that have come from these localities, no other ichnogenera have been found and the assemblage is unique in being so well sampled but of such low diversity, Indeed the assemblage completely lacks herbivore footprints. Herbivores appear within 100 ky after the Triassic-Jurassic boundary, in the form of both tracks and bones (e.g. prosauropods and small ornithischians) from other formations in eastern North America (Olsen et al., 2002). Apart from an apparent increase in the frequency of prosauropod tracks (*Otozoum*) and the size of the largest ornithischian tracks (*Anomoepus*), there is no obvious change in assemblage composition throughout the rest of the eastern North American record.

A quite remarkable feature of the assemblage from Stop 4 is that it records the appearance of the first truly large theropod dinosaurs, and increase in size (length) of over 20% that occurs at or just after the extinction level. A 20% increase in track length should scale to a doubling of mass, and thus this represents a very significant change in the top predators. As spelled out by us (Olsen et al., 2002), there are two obvious possible scenarios to explain the abrupt increase in size in theropod dinosaurs across the Triassic-Jurassic boundary in eastern North America (and globally). First, that the appearance of the much larger theropods represents a dispersal event from some unknown location, or second, that it represents an evolutionary event. We favor the second, evolutionary, hypothesis for the appearance of large theropods and suggest that the abrupt increase in size is most easily explained by a sudden evolutionary response of the theropod survivors (which may have been quite small) to ecological release, operating at time scales of thousands of years. We hypothesize a response similar to that inferred for reptiles on modern islands lacking competitors. This evolutionary response hypothesis could be falsified by the discovery of large theropod bones or diagnostic *Eubrontes giganteus* tracks in unquestionably Triassic strata. In any case, this track assemblage suggests that the dramatic drop in non-dinosaurian diversity was caused by an extrinsic environmental catastrophe, and the resulting drop in competitive pressure was the trigger for the global spread of large theropods.

Sauropodomorph and ornithischian dinosaurs also survived the Triassic-Jurassic boundary and joined theropods establishing the familiar dinosaurian-dominated ecological pattern that dominated the terrestrial world for the next 135 million years. The observations around the Triassic-Jurassic boundary are consistent with the hypothesis that the abrupt establishment of dinosaurian dominance was a response to massive ecological disruption and extinction of competitors, most likely caused by a giant bolide impact.

However, we also recognize that because of the very tight association with the overlying CAMP basalts we cannot rule out an indirect (via climate change) cause of the extinction. This scenario has been expressed by a number of workers, notably Hesselbo et al. (2002) and Beerling and Berner (2002). In this model, CO_2 from CAMP eruptions somewhere else at or preceding the Triassic-Jurassic boundary produce modest warming that triggers the release of vast amounts of methane clathrates creating a super-greenhouse and then upon oxidation produce a prolonged CO_2 super-greenhouse. Even the Ir anomaly could potentially be produce by a CAMP eruption.

The Orange Mountain basalt overlies the Passaic Formation and is a high-Ti quartz normative thoeliite (Puffer and Lechler, 1980) extremely widespread in the CAMP as the initial flow type. This flow is described in more detail at the following stop.

Stop 5: Orange Mountain and Preakness basalt of the CAMP, East and West Orange NJ, along I 280.

Major features:

1) Giant CAMP flood basalt flows

<u>Stop 5A: Orange Mountain Basalt. 40°47.483'N, 074°14.923'W.</u>

This huge cut in the Orange Mountain Basalt is type section of the formation (Olsen 1980a) (modified from Olsen and Schlische in Olsen et al. (1989). When this road was constructed in 1969, this 33 m high road cut was the deepest federally financed highway excavation east of the Mississippi River (Manspeizer, 1980).

A complete section of the lower flow unit of the Orange Mountain Basalt is exposed (Figure 22) (Olsen, 1980a,b; Manspeizer, 1980; Olsen et al., 1989). Unusual curved patterns of columnar joints are present in the cut, described as chevrons, oblique and reverse fans, and rosettes. This flow overlies red beds of the Passaic Formation that contains the Triassic-Jurassic boundary 10 m or so below the basalt contact. During the construction of this cut in 1969, PEO found several examples of the crocodiliomorph footprint *Batrachopus* sp. just beneath the contact with the Orange Mountain basalt.

At this outcrop of the type section, the west-dipping lower flow of the Orange Mountain Basalt is almost completely exposed (Olsen, 1980a; Puffer, 1987). The Orange Mountain Basalt at this exposure shows a complete, beautifully displayed Tomkeieff (1940) sequence (Olsen, 1980a; Puffer, 1987) almost exactly comparable to Long and Woods (1986), Type III flows. The thin (6 m) lower colonnade is fine-grained with large columns, the entablature is thick (35 m) with very well-developed curvicolumnar jointing, and the upper colonnade (10 m) is massive with poorly-developed columns.

According to Lyle (2000) the curvicolumnar jointing of the type seen at this outcrop is a direct result of ponded water on top of the cooling flows. The curved columns result from joints forming normal to large widely-spaced vertical master fractures in a hexagonal array that form very early in the cooling of the basalt flow. Water percolates down the master fractures and provides a surface cooling faster than adjacent basalt resulting in surface-normal hexagonal fractures that propagate away from the master fractures and curving towards the upper cooling colonnade of the flow. The master fractures can be clearly seen separating bowl-shaped fans of the curvicolumnar jointing. Additional bowl-shaped fractures, as seen here, may also have helped to control the radiating pattern of columns.

The Orange Mountain Basalt was fed by the Palisade sill (Ratcliffe, 1988) and it (and correlative units in other basins) represents the oldest of the known CAMP lava flows, flowing out less than 20 ky after the Triassic-Jurassic boundary (Olsen et al., 1996a; Olsen et al., 2002).

Stop 5B: Preakness Basalt. 40°47.946'N, 074°16.302'W.

This type section (Olsen, 1980a,b) exposes about 100 m of the lowest flow of the Preakness Basalt, the thickest extrusive multiple flow unit in the Newark basin section (Figure 23). It is a high-Ti, high-Fe quartz-normative tholeiite indistinguishable from the Holyoke Basalt of the Hartford basin and the Sander Basalt of the Culpeper basin. The lowest flow is characterized by a thin lower colonnade; a massive, thick, and coarse-grained entablature with very characteristic splintery columns; a massive upper colonnade; and a comparatively thin vesicular flow top. It does not closely resemble any of the flow types of Long and Wood (1986), but it is similar to the second flow of the Sander Basalt of the Culpeper basin, and the second flow of the Holyoke Basalt of the Hartford basin.

The splintery columns are defined by what Faust (1977) calls a platy prismatic joint system, which is characteristic of the lower flow throughout the areal extent of the Preakness Basalt. The joint system is not hexagonal and does not appear to be a faulting phenomena (although faulting does exaggerate it). Its origin remains a mystery.

At this locality the contact (now overgrown) with the underlying Feltville Formation is simple, with massive basalt of the lower colonnade in direct contact with sedimentary rocks. At other localities thin flow units, pillow lavas or rubble flows are present (Olsen, 1980a).

End of field trip.

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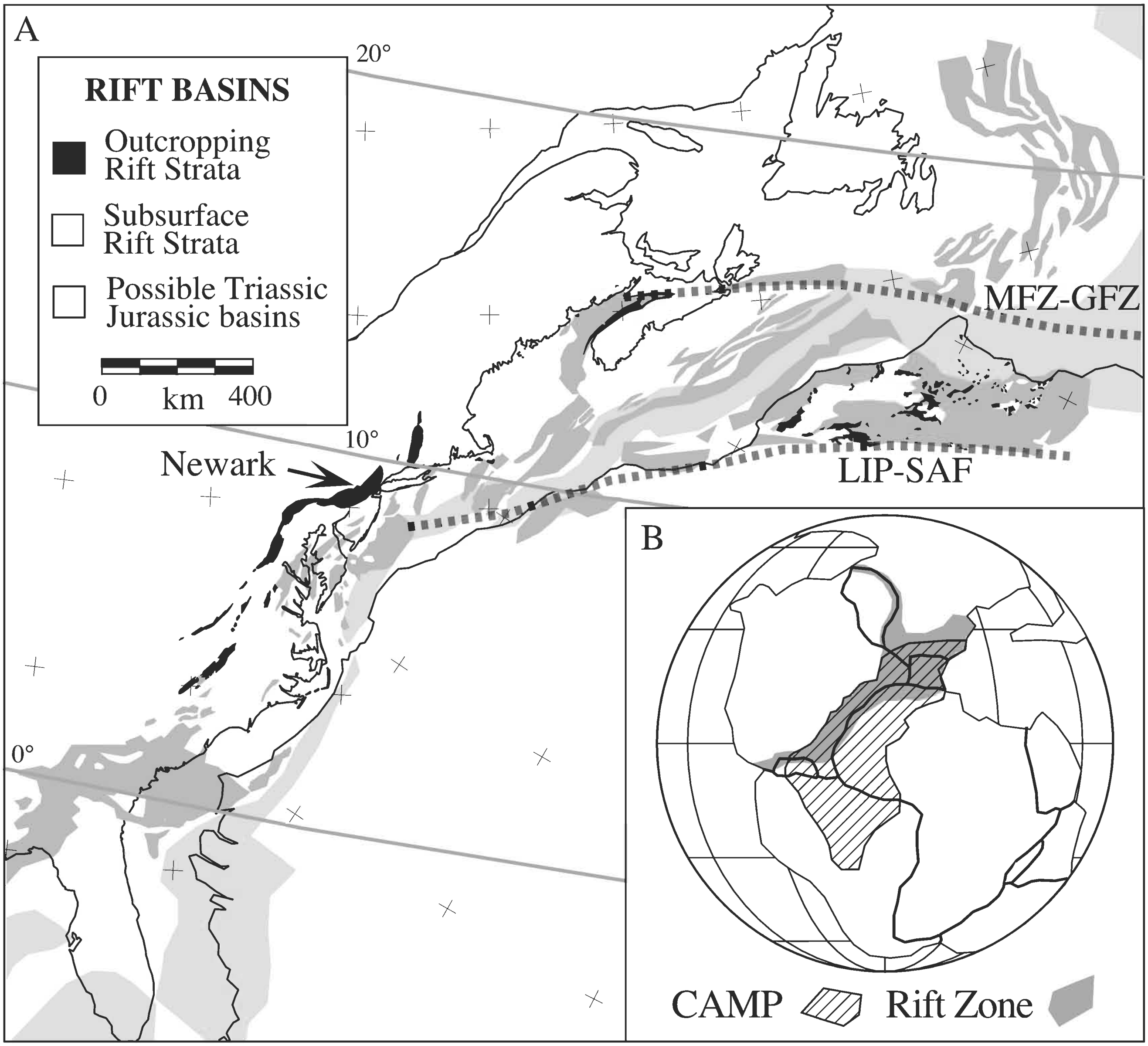


Figure 1. Central Pangean rift basins in the Early Jurassic Pangea relationship to the CAMP. **A**, rift basins of central Pangea showing the position of the Newark basin, the Long Island Platform - South Atlas fault zone (LIP-SAF) and the Minas - Gibraltar fault zone (MFZ-GFZ) (modified from Olsen, 1997). **B**, Pangea in the Early Jurassic showing the spatial relationship between the CAMP and the central Pangean rift zone.

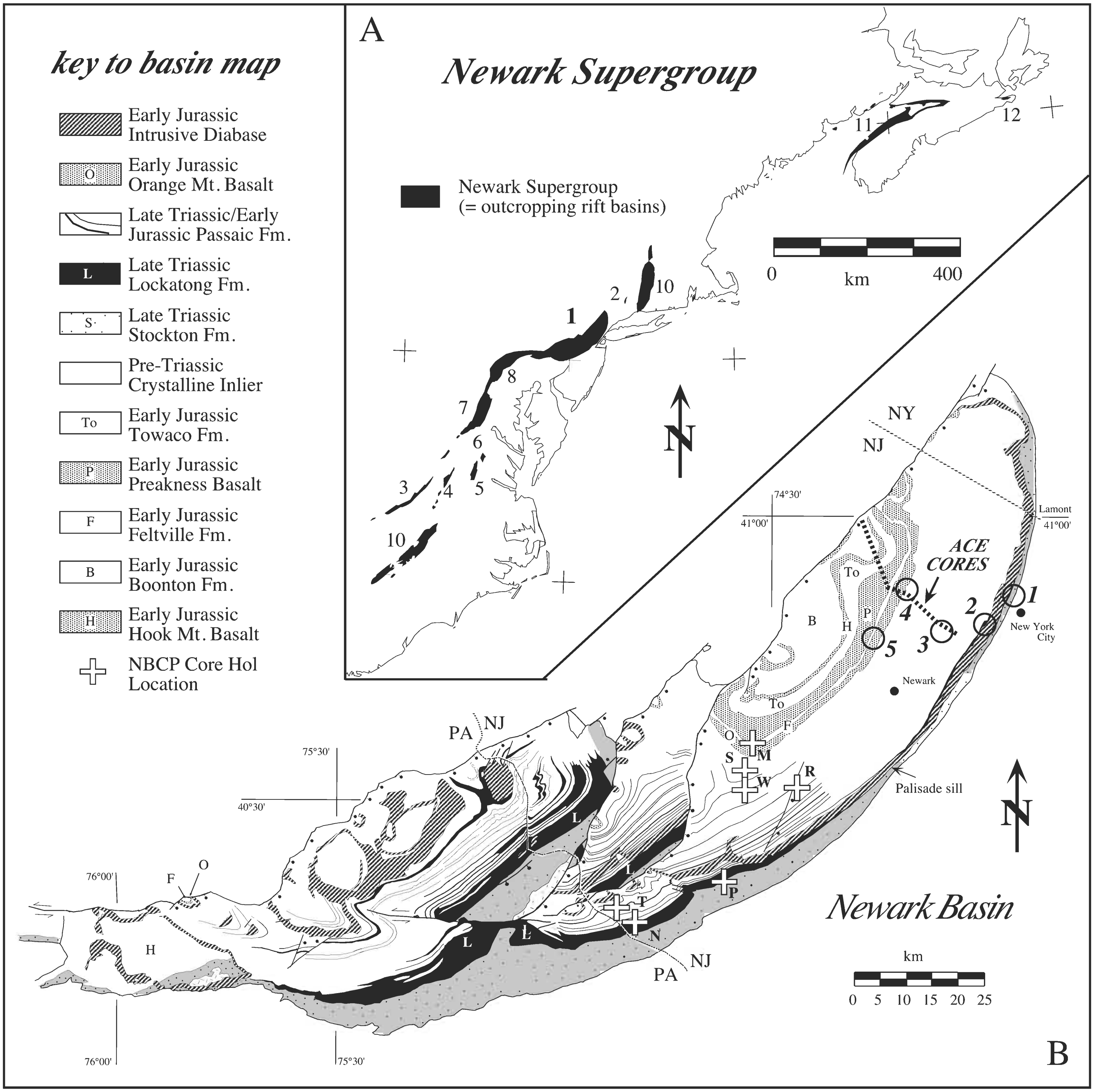
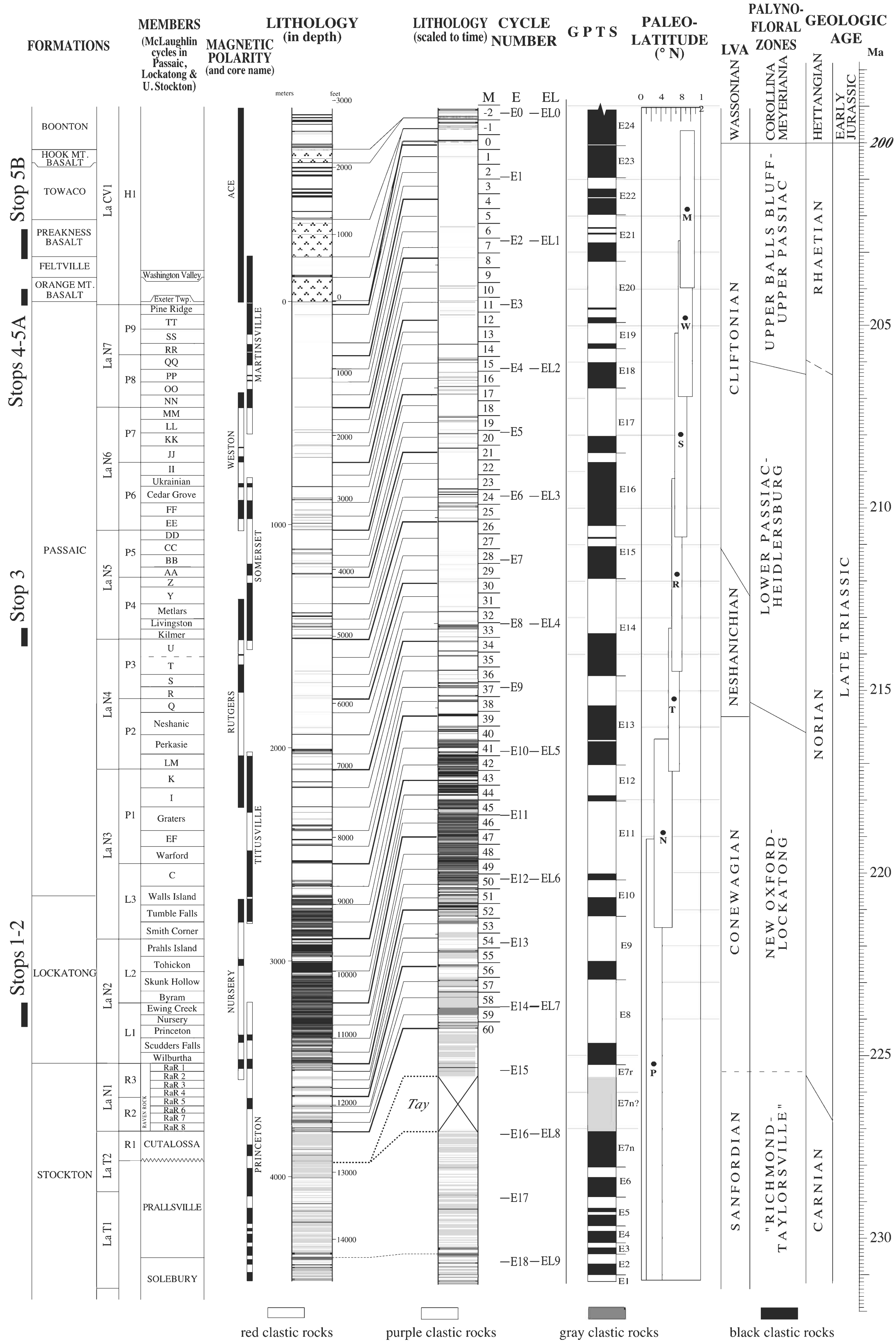


Figure 2. The Newark basin within the Newark Supergroup. **A**, Newark Supergroup: **1**, Newark basin; **2**, Hartford (below) and Deerfield (above) basins; **3**, Dan River basin; **4**, Farmville and associated basins; **5**, Richmond; **6**, mostly buried Taylorsville basins; **7**, Culpeper basin; **8**, Gettysburg basin; **9**, Pomperaug basin; **10**, Deep River basin; **11**, Fundy basin; **12**, mostly buried Chedabucto or Orpheus basin (modified from Olsen, 1997). **B**, Newark basin and lithostratigraphic divisions (modified from Olsen et al., 1996a). Stops (circles) are: ***1***, Stop 1 - Palisades Intserstate Park; ***2***, Stop 2 - Granton Quarry; ***3***, Stop 3 - Lyndhurst; ***4***, Stop 4 - Tilcon Clifton Quarry; ***5***, Stop 5 - Rt 280 road cuts.

Figure 3. Newark and Hartford basin section and combined time scale showing distribution of field stops (adapted from Olsen and Kent, 1999; Olsen et al., 2002).



FORMATIONS
MEMBERS (McLaughlin cycles in Passaic, Lockatong & U. Stockton)
MAGNETIC POLARITY (and core name)
LITHOLOGY (in depth)
LITHOLOGY (scaled to time)
CYCLE NUMBER
G P T S
PALEO-LATITUDE (° N)
LVA
PALYNO-FLORAL ZONES
GEOLOGIC AGE
Ma
Stop 5B
Stops 4-5A
Stop 3
Stops 1-2
BOONTON
HOOK MT. BASALT
TOWACO
PREAKNESS BASALT
FELTVILLE
ORANGE MT. BASALT
PASSAIC
LOCKATONG
STOCKTON
La CV1
La N7
La N6
La N5
La N4
La N3
La N2
La N1
La T2
La T1
H1
P9
P8
P7
P6
P5
P4
P3
P2
P1
L3
L2
L1
R3
R2
R1
Washington Valley
Exeter Twp.
Pine Ridge
TT
SS
RR
QQ
PP
OO
NN
MM
LL
KK
JJ
II
Ukrainian
Cedar Grove
FF
EE
DD
CC
BB
AA
Z
Y
Metlars
Livingston
Kilmer
U
T
S
R
Q
Neshanic
Perkasie
LM
K
I
Graters
EF
Warford
C
Walls Island
Tumble Falls
Smith Corner
Prahls Island
Tohickon
Skunk Hollow
Byram
Ewing Creek
Nursery
Princeton
Scudders Falls
Wilburtha
RAVEN ROCK
RaR 1
RaR 2
RaR 3
RaR 4
RaR 5
RaR 6
RaR 7
RaR 8
CUTALOSSA
PRALLSVILLE
SOLEBURY
ACE
MARTINSVILLE
WESTON
SOMERSET
RUTGERS
TITUSVILLE
NURSERY
PRINCETON
meters
feet
Tay
M E EL
E0 EL0
E1
E2 EL1
E3
E4 EL2
E5
E6 EL3
E7
E8 EL4
E9
E10 EL5
E11
E12 EL6
E13
E14 EL7
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E7n
E6
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E3
E2
E1
0 4 8 12
M
W
S
R
T
N
P
WASSONIAN
CLIFTONIAN
NESHANICHIAN
CONEWAGIAN
SANFORDIAN
COROLLINA MEYERIANIA
UPPER BALLS BLUFF-UPPER PASSAIC
LOWER PASSAIC-HEIDLERSBURG
NEW OXFORD-LOCKATONG
"RICHMOND-TAYLORSVILLE"
HETTANGIAN
RHAETIAN
NORIAN
CARNIAN
EARLY JURASSIC
LATE TRIASSIC
200
205
210
215
220
225
230
red clastic rocks
purple clastic rocks
gray clastic rocks
black clastic rocks

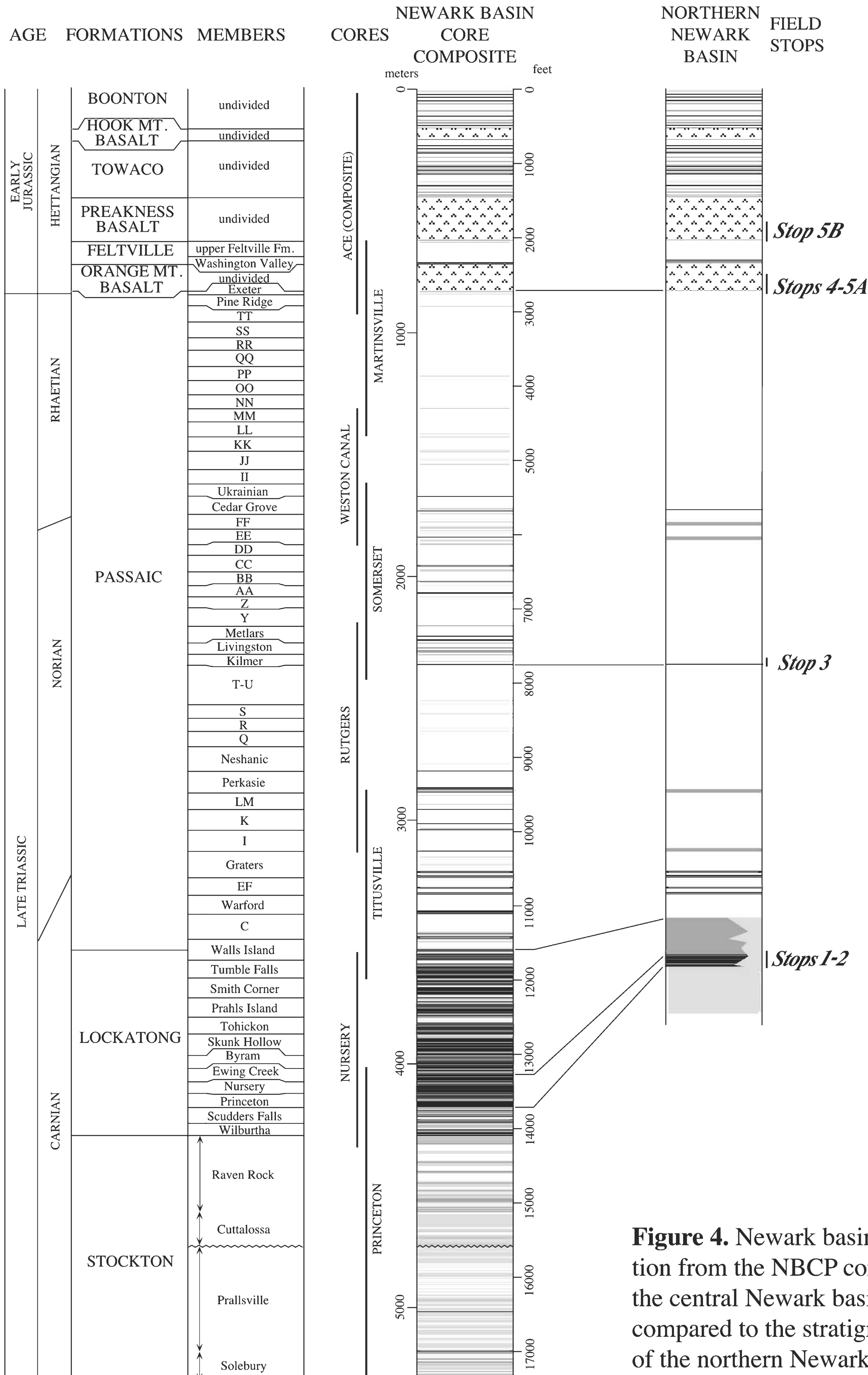
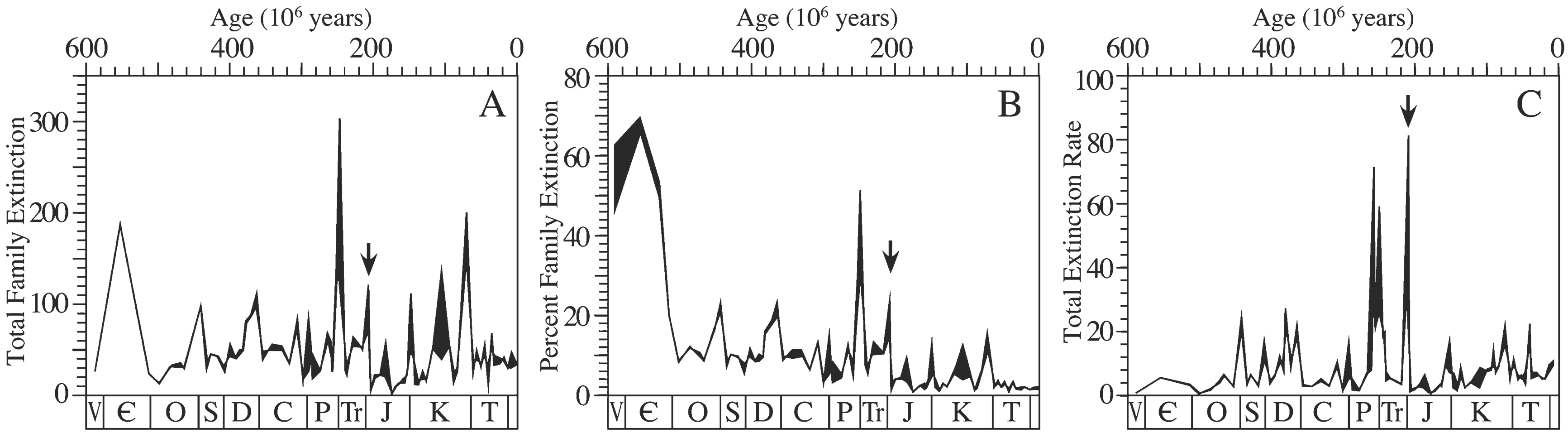


Figure 4. Newark basin section from the NBCP cores of the central Newark basin (left) compared to the stratigraphy of the northern Newark basin with field trips stops. See Figure 3 for key to lithologies.



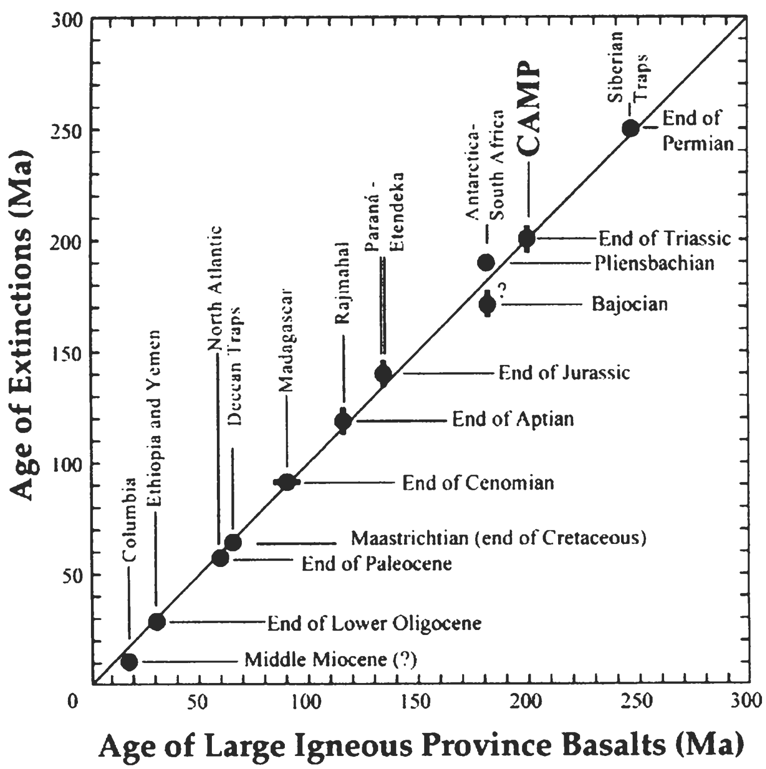


Figure 5. Above. Extinction rate of marine and continental organisms through the last 600 million years (from Benton, 1995) with arrow at Triassic-Jurassic boundary. The upper and lower bounds represent the maximum and minimum curves. A, Extinction rate expressed as the numbers of families that died out in each stratigraphic stage. B, Extinction rate expressed as a percent of families that died out in relation to contemporaneous diversity in each stratigraphic stage. C, Extinction rate expressed as the number of families that died out in relation to the duration of each stratigraphic stage.

Figure 6. Left. Correlation between the timing for eruption of various large igneous provinces to the timing of mass extinction events over the past 250 m.y. From Hames et al. (2003), modified from Courtillot (1999).

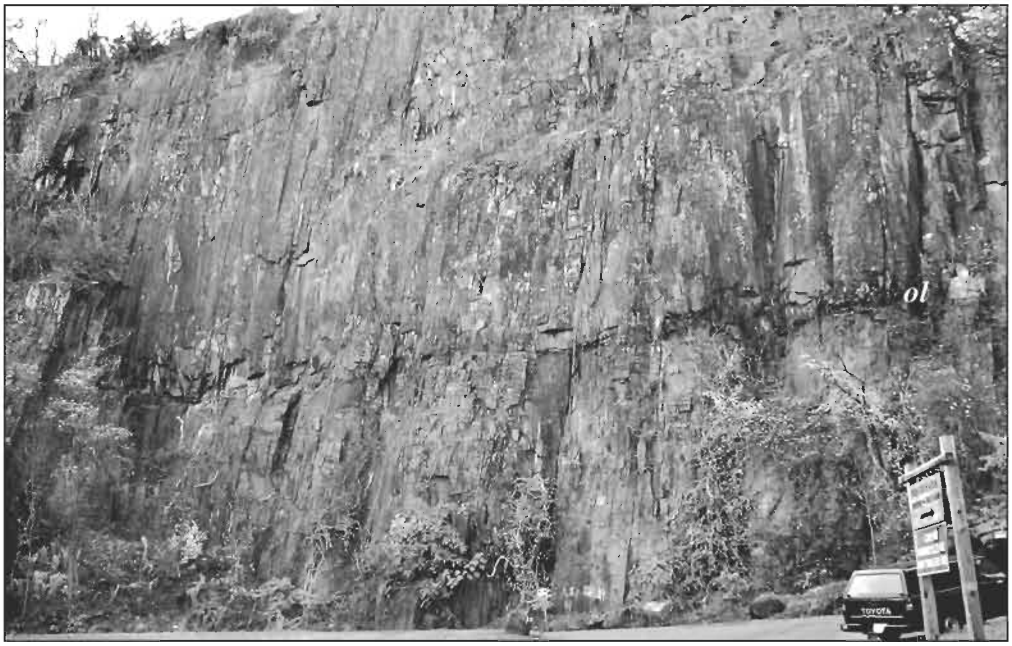


Figure 7. Quarry face at Stop 1A, with olivine zone (ol) in zone of deflected columns.

Figure 8. Mostly concordant contact of Palisade sill with mudstone and arkose of Lockatong Formation just north of the George Washington Bridge at Stop 1b

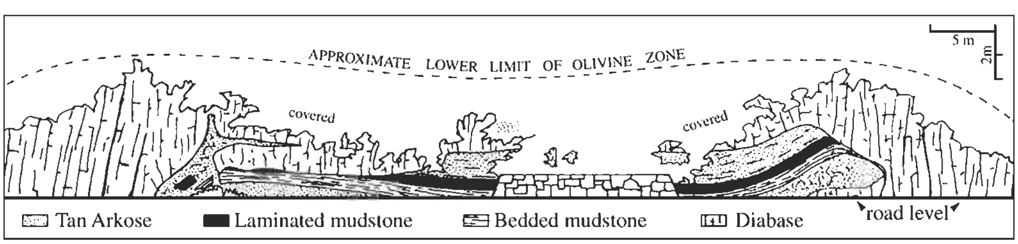


Figure 9. Sketch of discordant contact of Palisade sill and Lockatong Formation along Henry Hudson Drive just south of the George Washington Bridge at Stop 1c. (from Olsen, 1980a).

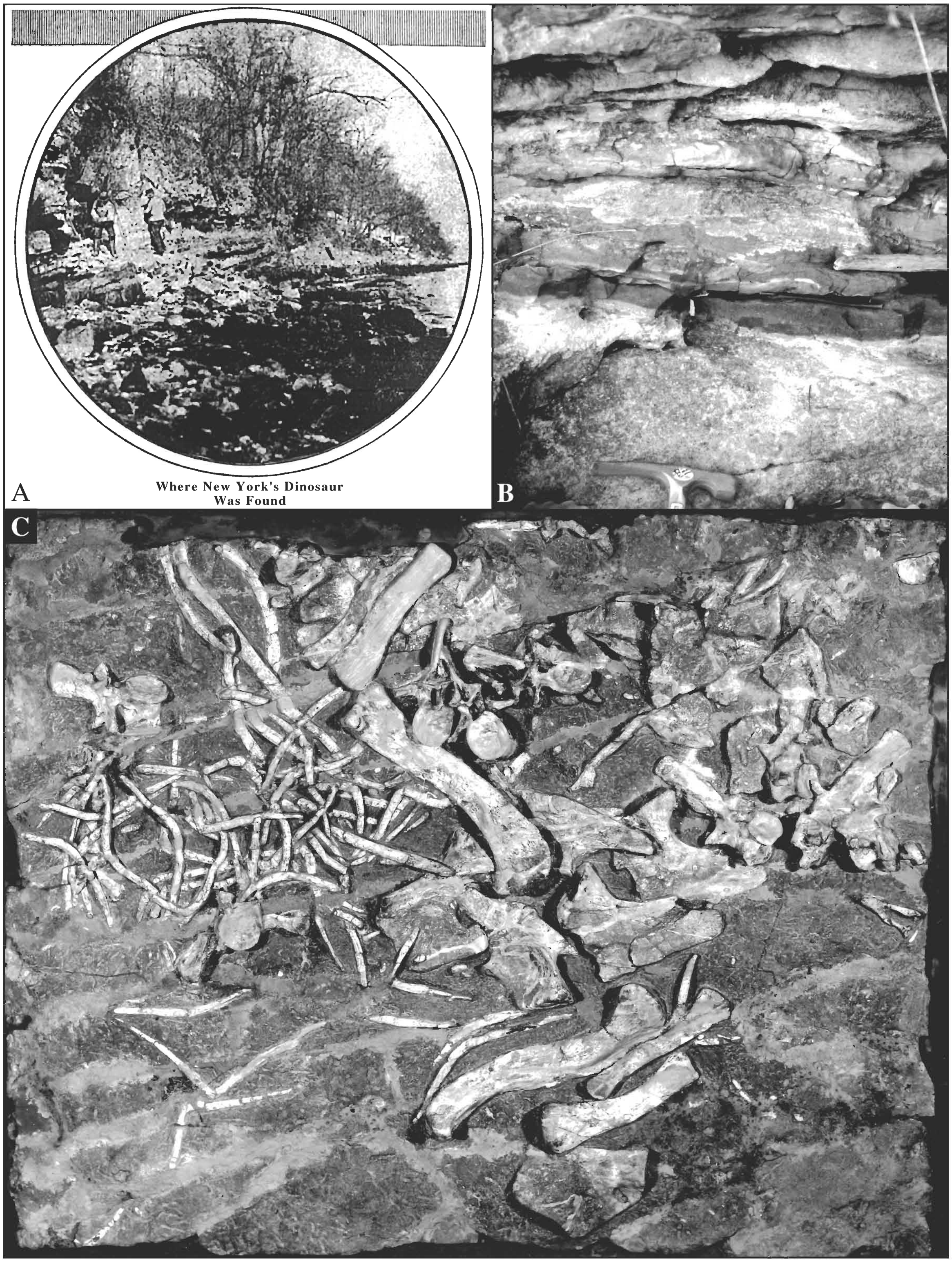


Figure 10. *Rutiodon mahattanensis* and outcrops at Stop 1d, Stockton Formation: **A**, photograph from the front page of the magazine section of the New York Times for December 25, 1910, showing the location of the phytosaur skeleton just south of the boundary with the Palisades Interstate Park (Stop 3g) (with permission of the New York Times); **B**, photograph of typical lithologies (purple and red mudstones and tan arkose) at the north end of the outcrops shown in A; **C**, disarticulated partial skeleton of the large phytosaur *Rutiodon mahattanensis* (AMNH 4991) (courtesy of the American Museum of Natural History).

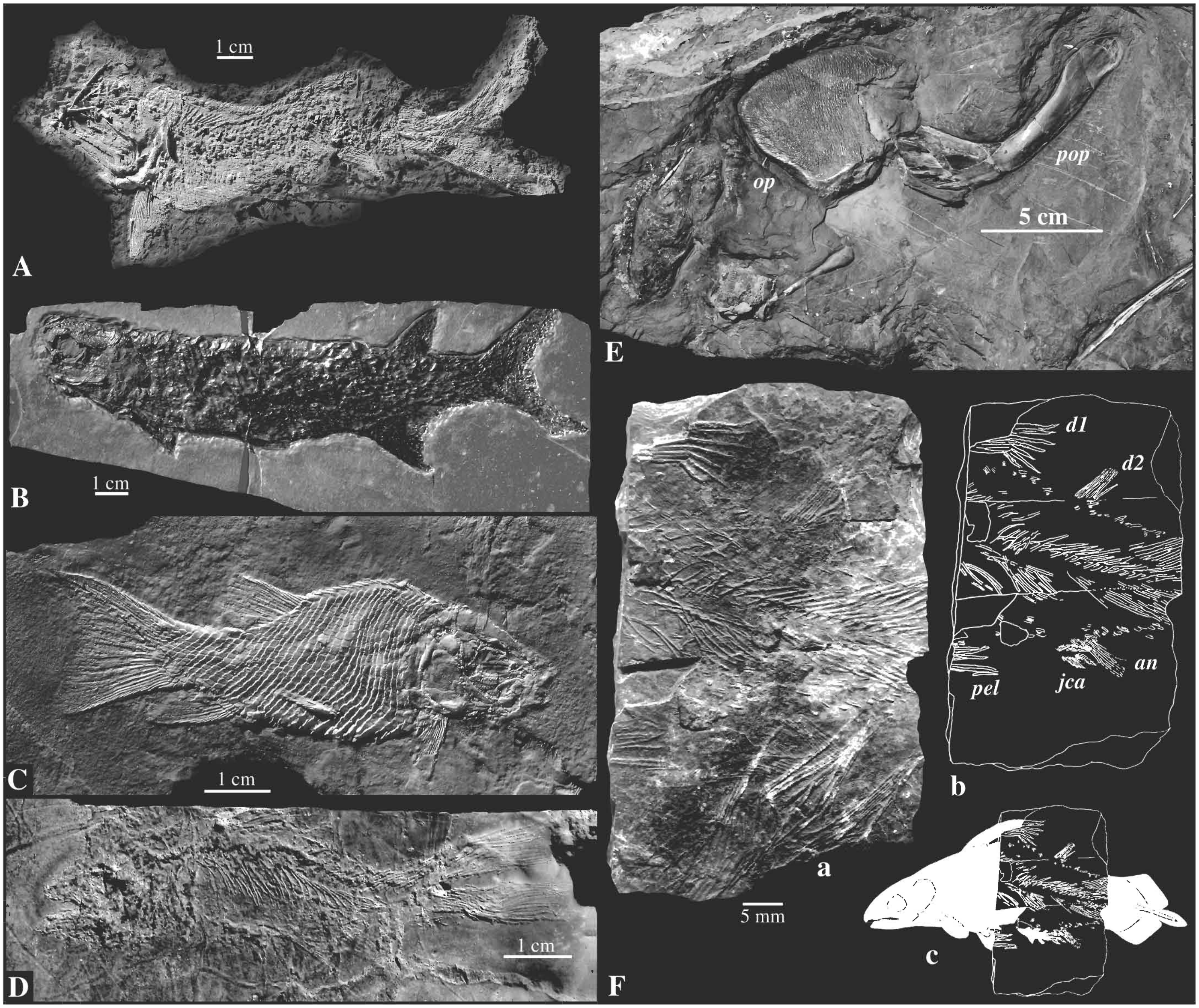
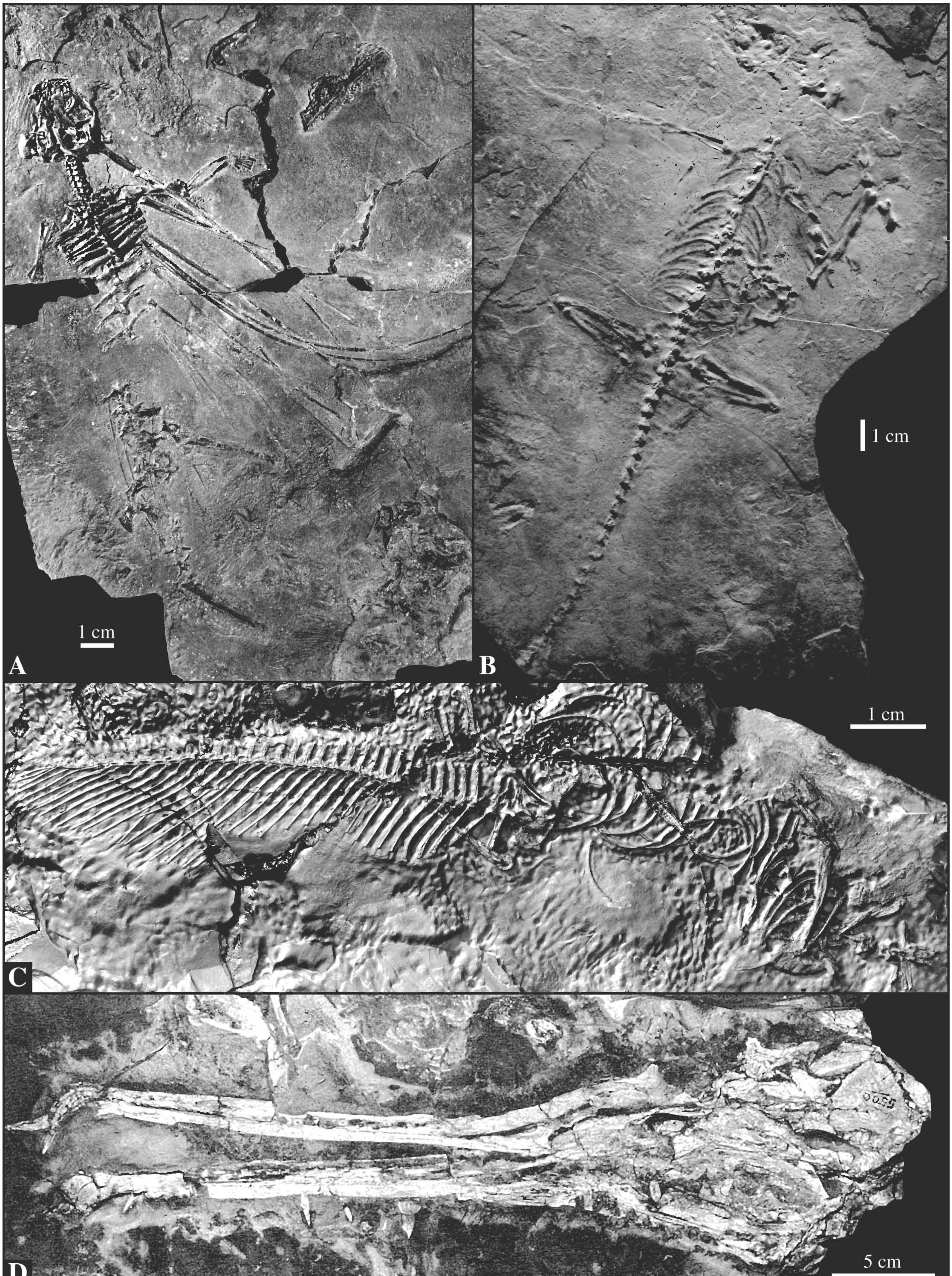


Figure 11A. Vertebrates of the Lockatong Formation for Stop 2 (from Olsen and Rainforth, 2003). Examples of fish from Lockatong Formation. A, polysulfide cast of the palaeoniscoid *Turseodus* sp. from cycle W-5, Yale Weehawken quarry (YPM field collection number W5-663). B, the refieldiid palaeonisciform *Synorichtheys* sp. from Gwynedd, Pennsylvania (YPM 8863). C, the holostean *Semionotus braunii*, from cycle W-6, W-5, Yale Weehawken quarry. D, the small coelacanth Osteopleurus newarki from cycle G-7, Granton Quarry (Stop 2). E, disarticulated partial skull of the large coelacanth cf. *Pariostegeus* sp. from float at Granton Quarry (Stop 2) (NJSM 16697): pop, preopercular; op, opercular. F, fragmentary, articulated specimen of *Osteopleurus newarki* from float at Granton Quarry (Stop 2) that seems to have been buried in the process of giving birth to a baby (see Rizzo, 1999) (NJSM 15819): a, slab preserving mid-section of probable adult and juvenile; b, drawing of same (d1, anterior dorsal fin; d2, posterior dorsal fin; pcl, pelvic fins; an, anal fin; jca, juvenile (caudal fin) in position of cloaca, just in front of anal fin); c, orientation of slab relative to outline of complete fish.

Figure 11B: Reptiles from the Lockatong Formation at Granton Quarry (Stop 2): **A**, type specimen of *Icarosaurus seifkeri* from cycle G-3 (uncataloged AMNH specimen) (from Colbert, 1966; with permission of the American Museum of Natural History); **B**, female *Tanytrachelos ahynis* found by Steven Stelz, Trinny Stelz and James Leonard; **C**, type specimen of *Hypuronector limnaios* from cycle G7 (AMNH 7759) (from Colbert and Olsen, 2001; with permission of the American Museum of Natural History); **D**, skull of cf. *Rutiodon carolinensis* found in float (AMNH 5500) (from Colbert, 1965; with permission of the American Museum of Natural History).


A
1 cm
B
1 cm
C
1 cm
D
5 cm

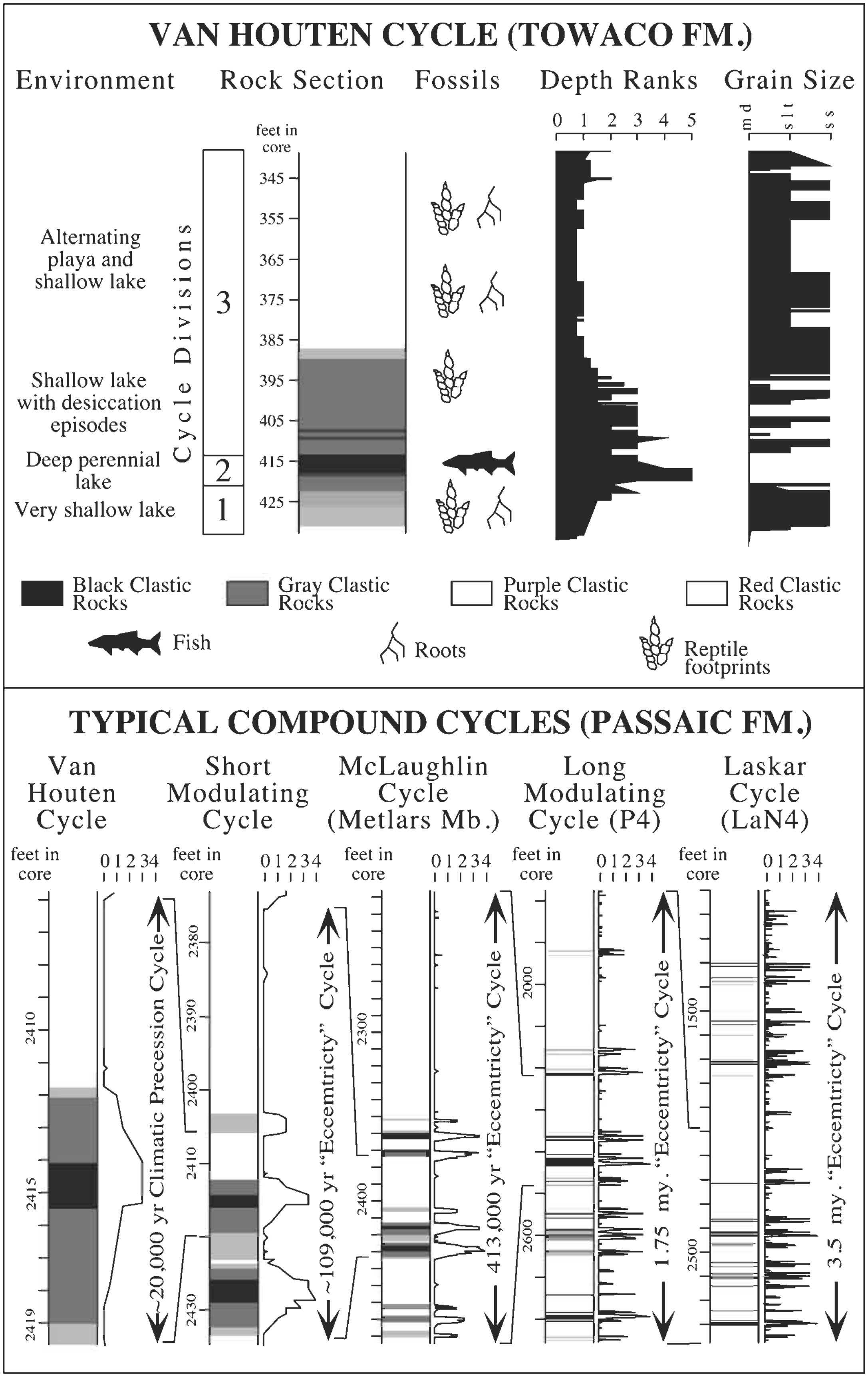


Figure 12. Van Houten and compound cycles (modified from Olsen and Kent, 1999).

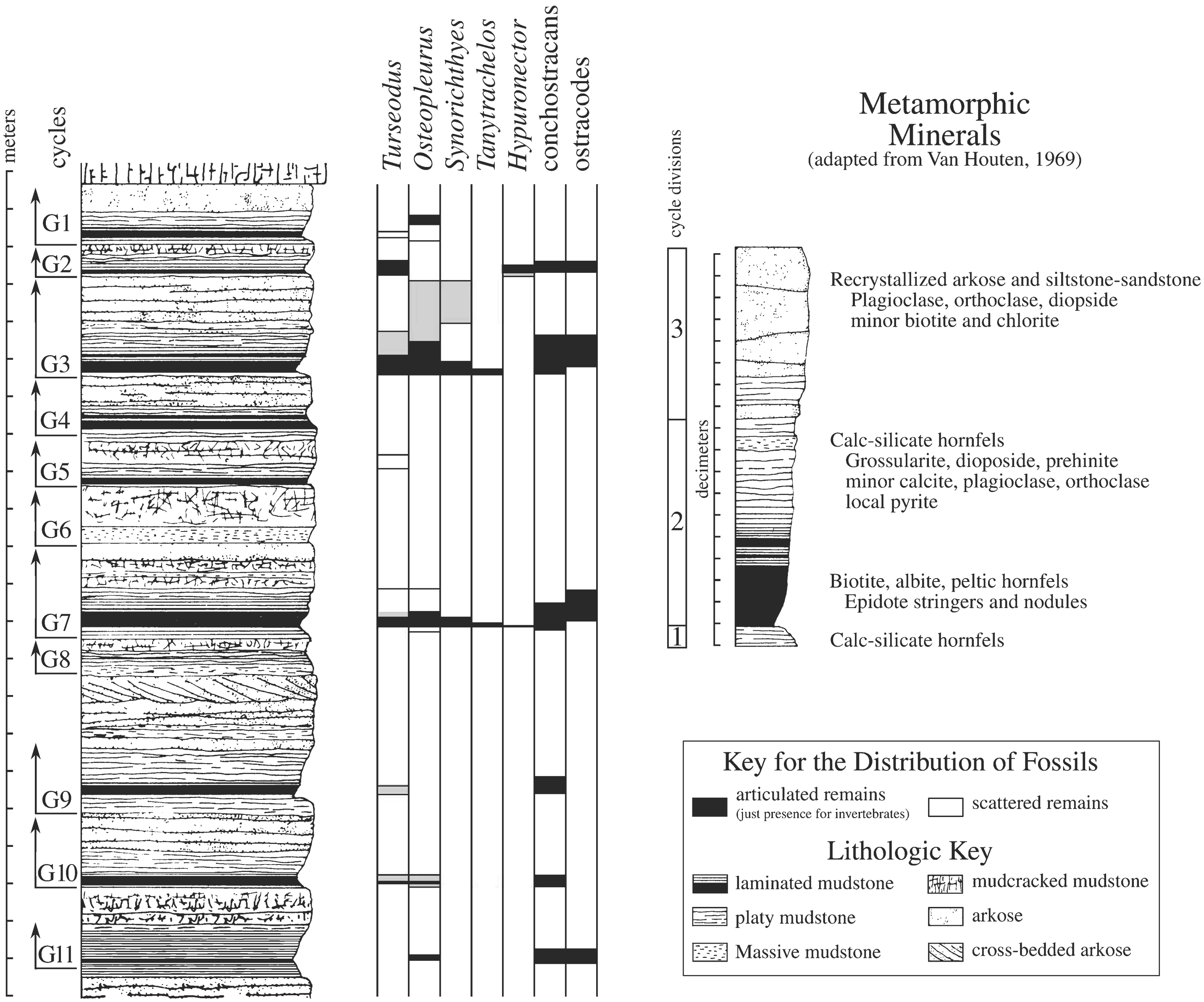


Figure 13. Section and fossils at Granton Quarry, Stop 2 (modified from Olsen, 1980b).

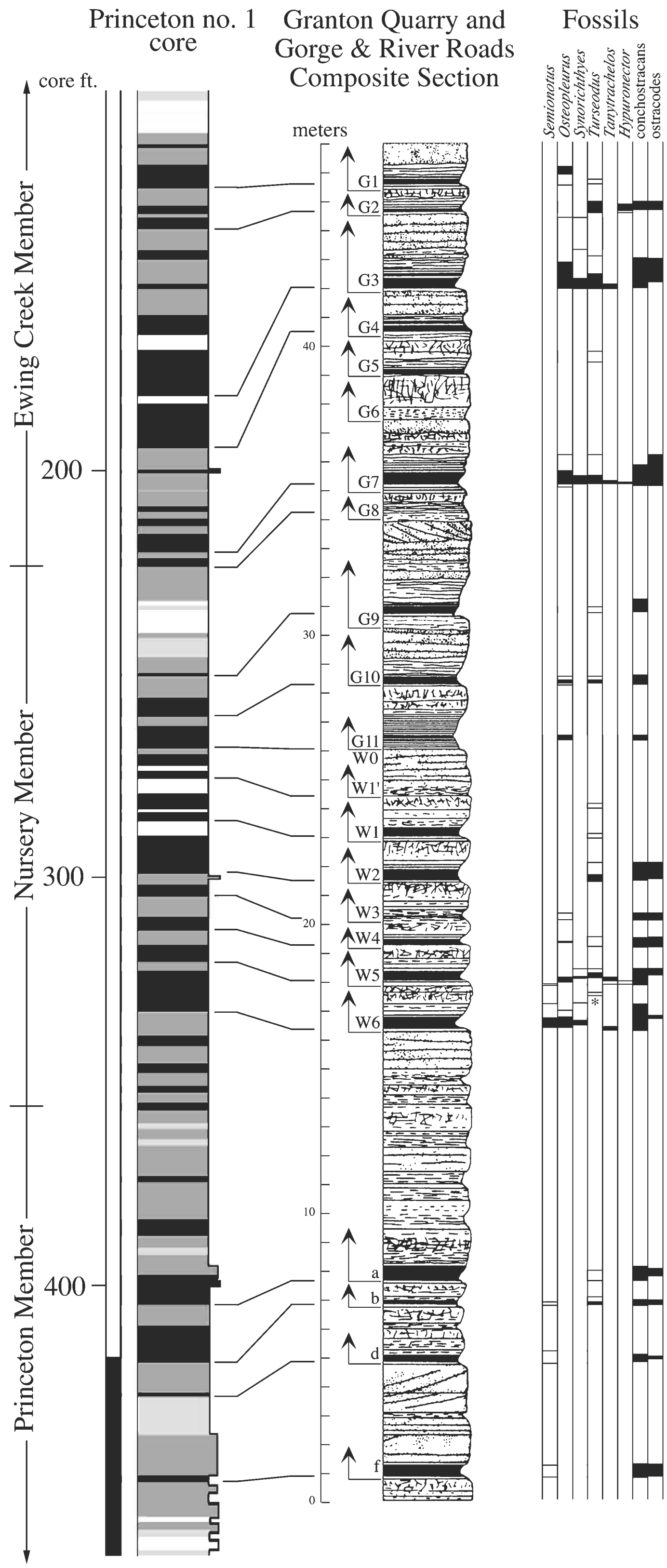


Figure 14. Composite section of the Lockatong in vicinity of Stops 1 and 2 with the distribution of fossils and correlation to Newark basin coring project core, Princeton no. 1. Key for lithologies of the Princeton no. 1 core in Figure 2 and for northeast composite in Figure 13.

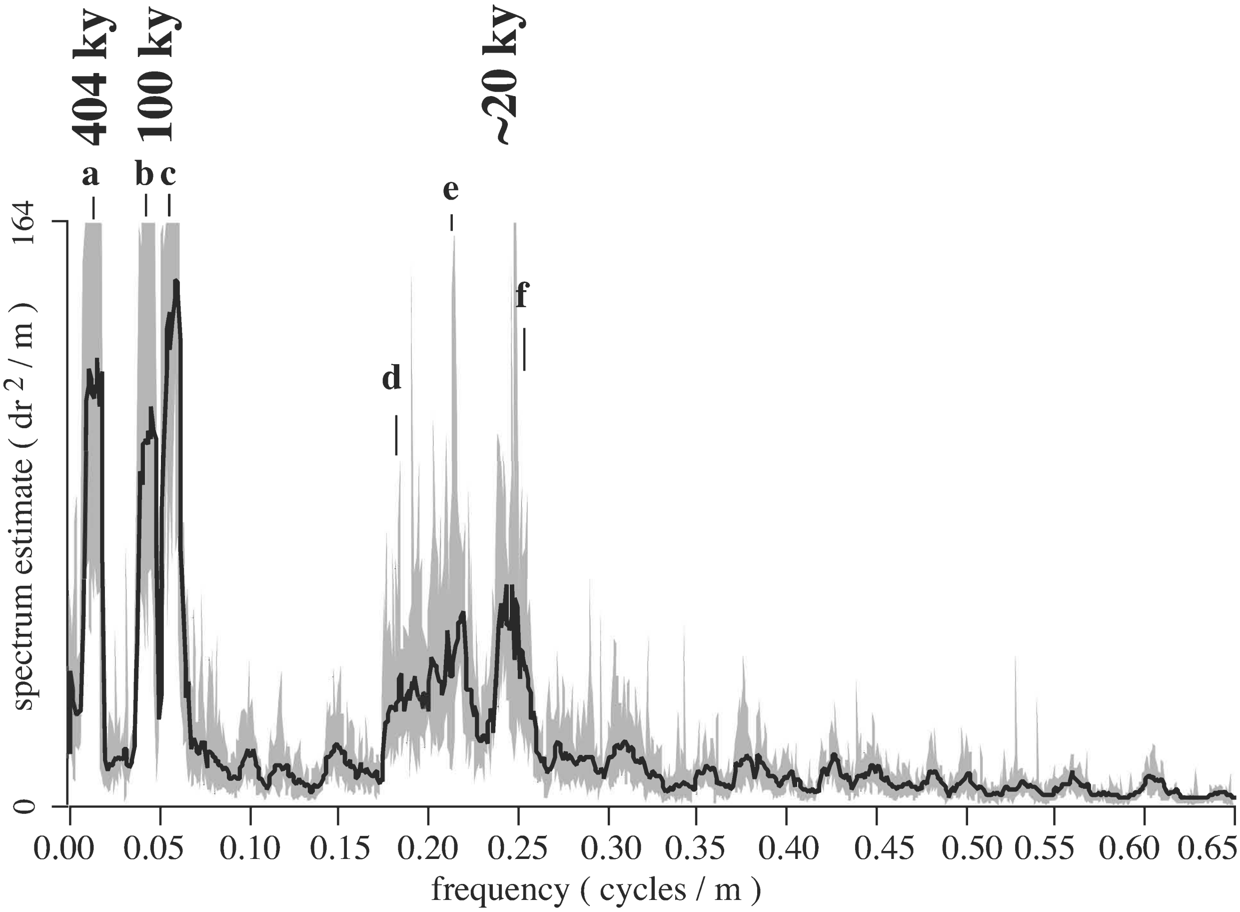


Figure 15. Power spectrum of Lockatong and lowest Passaic Formations in the NBCP cores (from Olsen and Kent, 1996). Multiple-window spectrum estimate of periodicties. Shaded zone shows 90 % confidence intervals (upper limit locally truncated). Spectrum computed using half bandwidth (NW) of 3.5, 6 orders of eigenspectra, _t = 1 ft, and with the time scale set by assuming that spectral peaks a is 404 ky cycle. Specific spectral peaks are: a, 73.9 m, 404 ky; b, 22.8 m 124.7 ky; c, 17.4 m, 95.1; d, 5.5 m, 30.0; e, 4.6 m. 25.2 ky; f, 3.9 m, 21.3 ky.

Figure 16. Exposures of copper-bearing white and tan sandstone surrounding purple siltstone of lower Kilmer Member and upper member T-U at Stop 3, Lyndhurst, New Jersey.

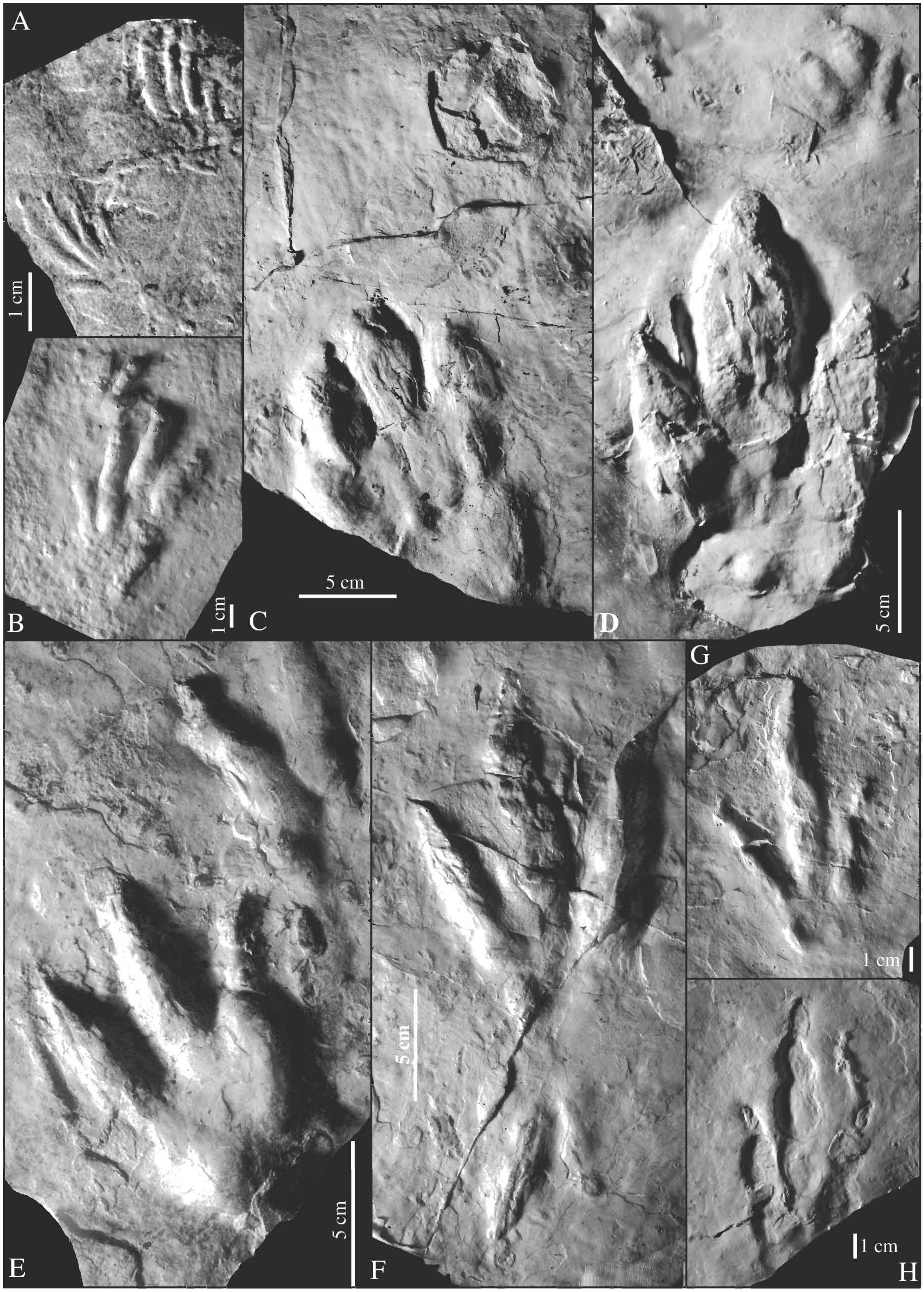


Figure 17. Footprints from Stop 3 at Lyndhurst: A, lepidosauromorph track *Rhynchosauroides* sp; B, probable rauisuchian track *Brachychirotherium* sp.; C, probable rauisuchian track *Brachychirotherium parvum*; D, ornithischian dinosaurian track *Atreipus milfordensis*; E, ?saurischian dinosaurian track "new genus 1"; F, theropod dinosaur tracks *Anchisauripus* sp. (above) and *Grallator* sp. (below); G, theropod dinosaur track *Grallator* sp.; H, theropod dinosaur track *Grallator* cf. *G. parallelus*.

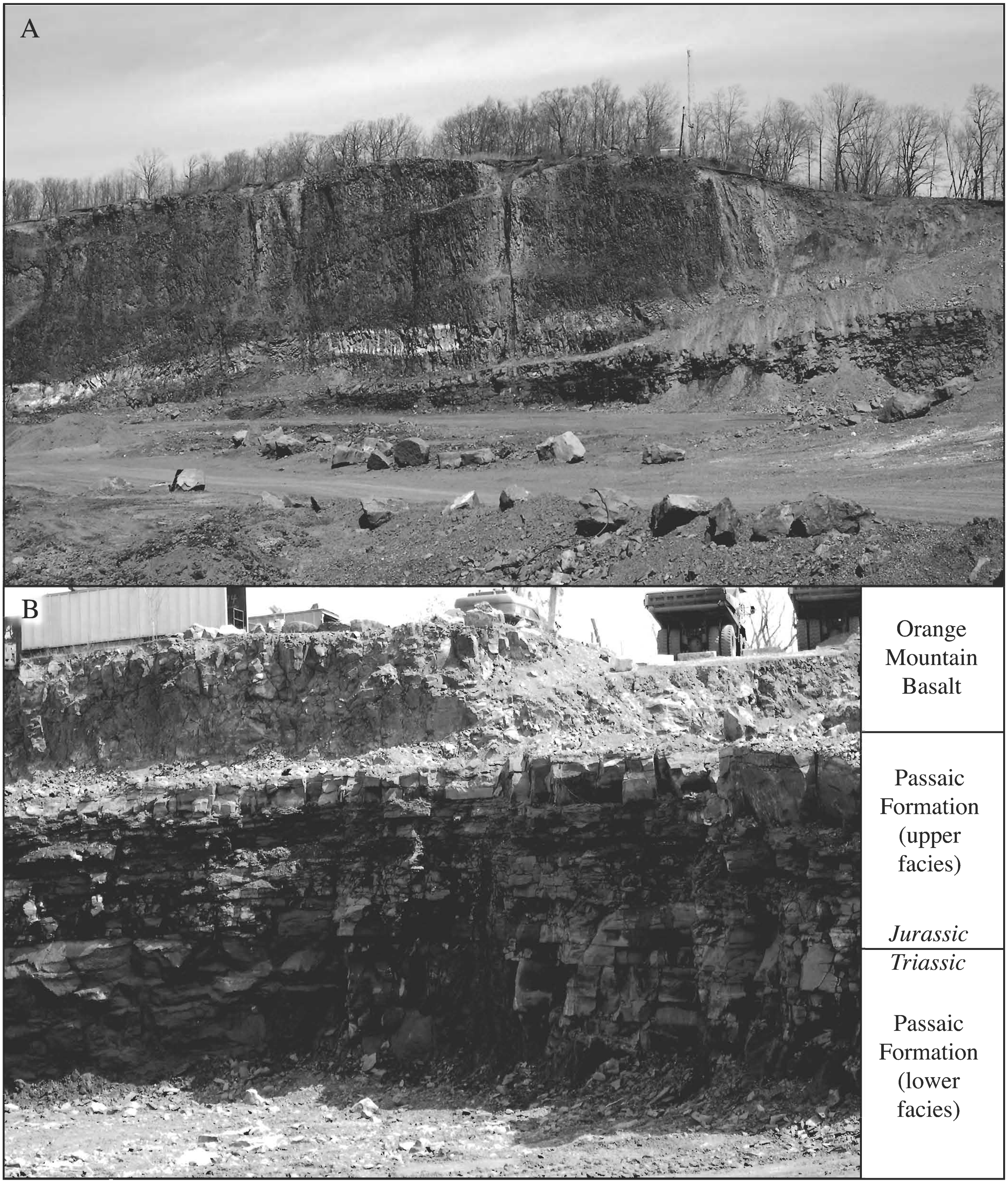


Figure 18. Stop 4: A, North wall of Tilcon Clifton Quarry showing the lower flow of the Orange Mountain Basalt resting on Passaic Formation containing the Triassic-Jurassic boundary. B, Section on east face of quarry showing basic straigraphy and possible minor unconformity at Triassic-Jurassic boundary.

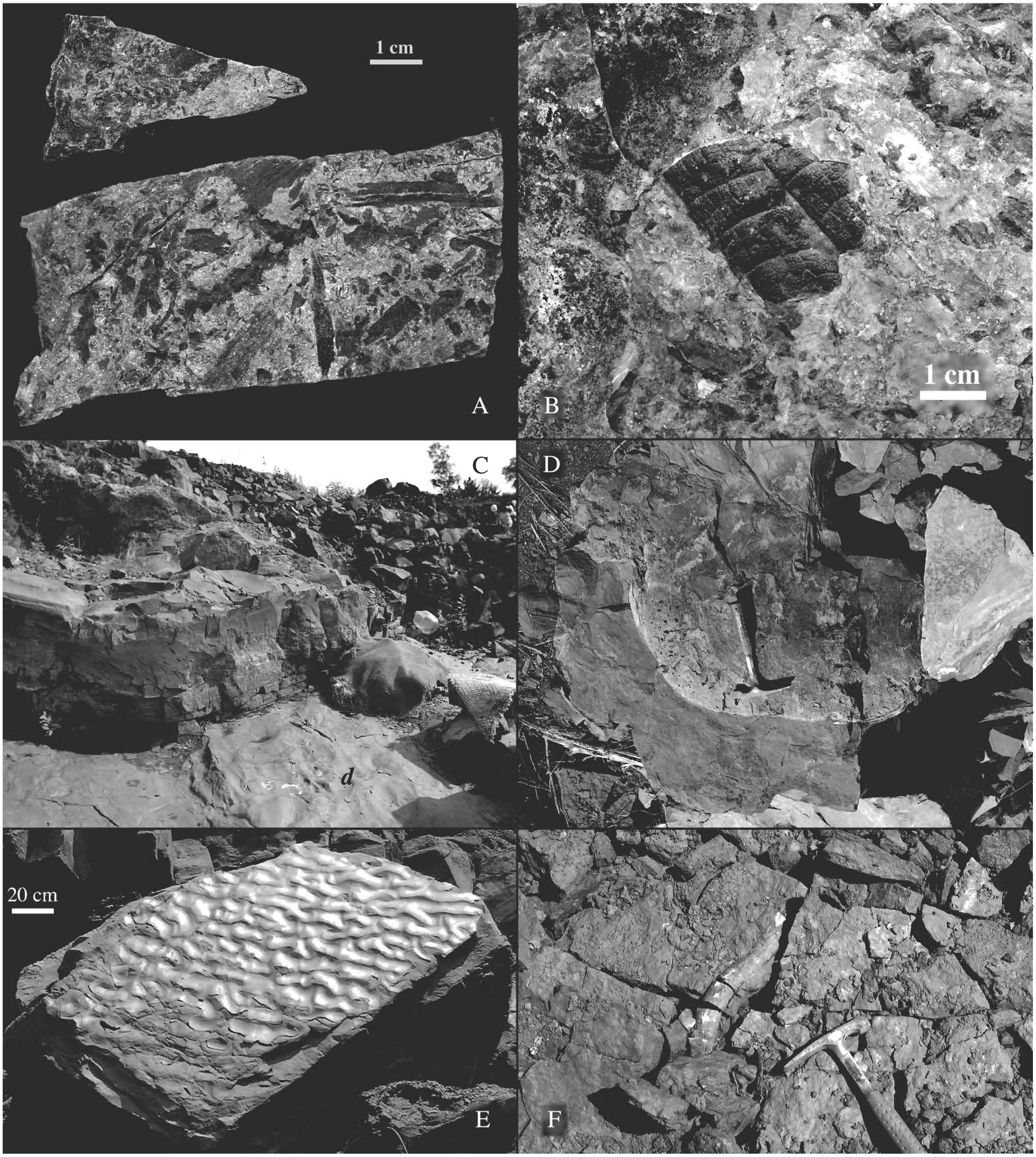


Figure 19.Fossils and structures from Stop 4: A, Shoots of the conifer *Brachyphyllum*; B, fragment of the dipteridaceous fern *Clathropteris meniscoides*; C, contact between Passaic Formation and overlying Orange Mountain Basalt (***d*** is an in situ dinosaur track) – note tilted pipe vesicles; D, contact between Passaic Formation and overlying Orange Mountain Basalt showing filling of a channel by basalt – the contact cuts across bedding; E, rippled surface of ripple-cross-laminated fine sandstone block, upper facies; F, casts of roots and possibly stems in red mudstone with patches of organic preserved plants.

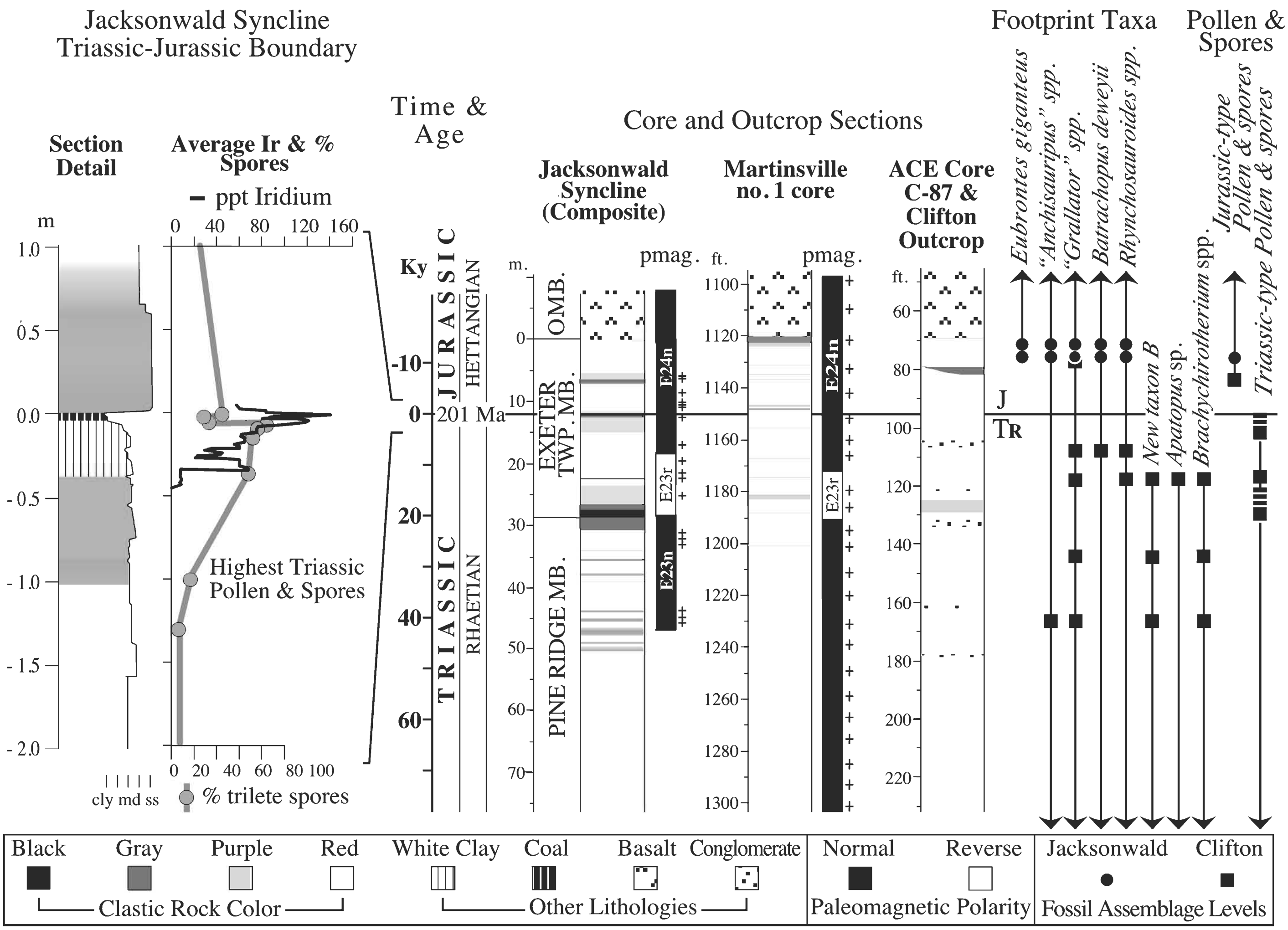


Figure 20. Fine-scale correlation of the Triassic-Jurassic boundary between outcrop, the Martinsville no. 1 core of the NBCP, and ACE core PT-87, and associated biotic and geochemical changes (from Olsen et al., 2002).

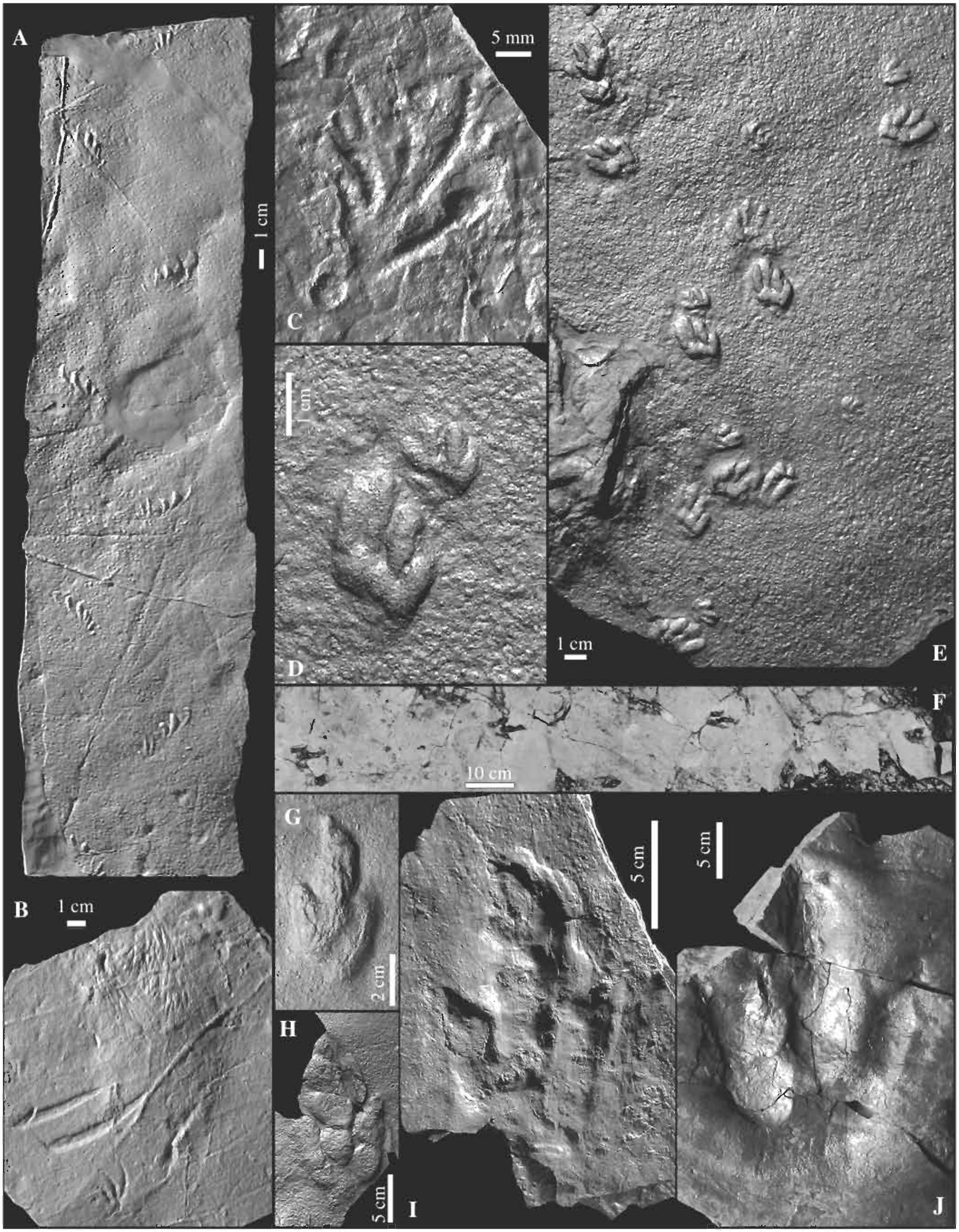


Figure 21. Earliest Jurassic age reptile footprints from the uppermost Passaic Formation of the area around Stop 4: A-C, *Rhynchosauroides* n. sp.; D-E, *Batrachopus deweyii*; F, trackway of medium-sized *Anchisauripus sillimani*; G, large *Grallator* sp.; H, medium-sized *Anchisauripus* sp.; I, *Grallator tuu berosus*; J, *Eubrontes giganteus*.

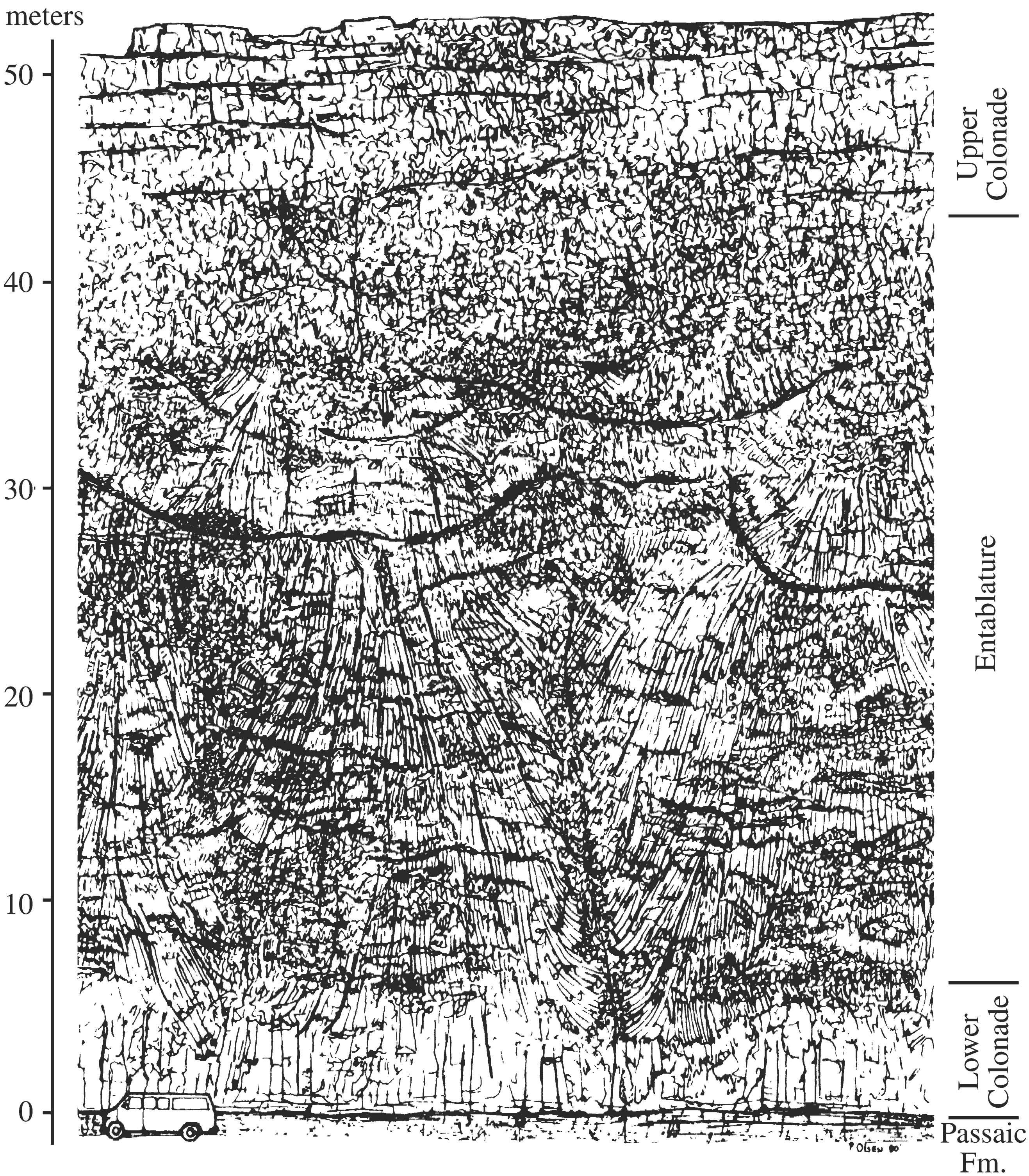


Figure 22. Composite drawing of the type section of the Orange Mountain Basalt along Interstate Route 280 in East Orange, New Jersey. Traced from a composite of a continuous series of photographs with the dip removed and compiled vertically (Olsen, 1980b).

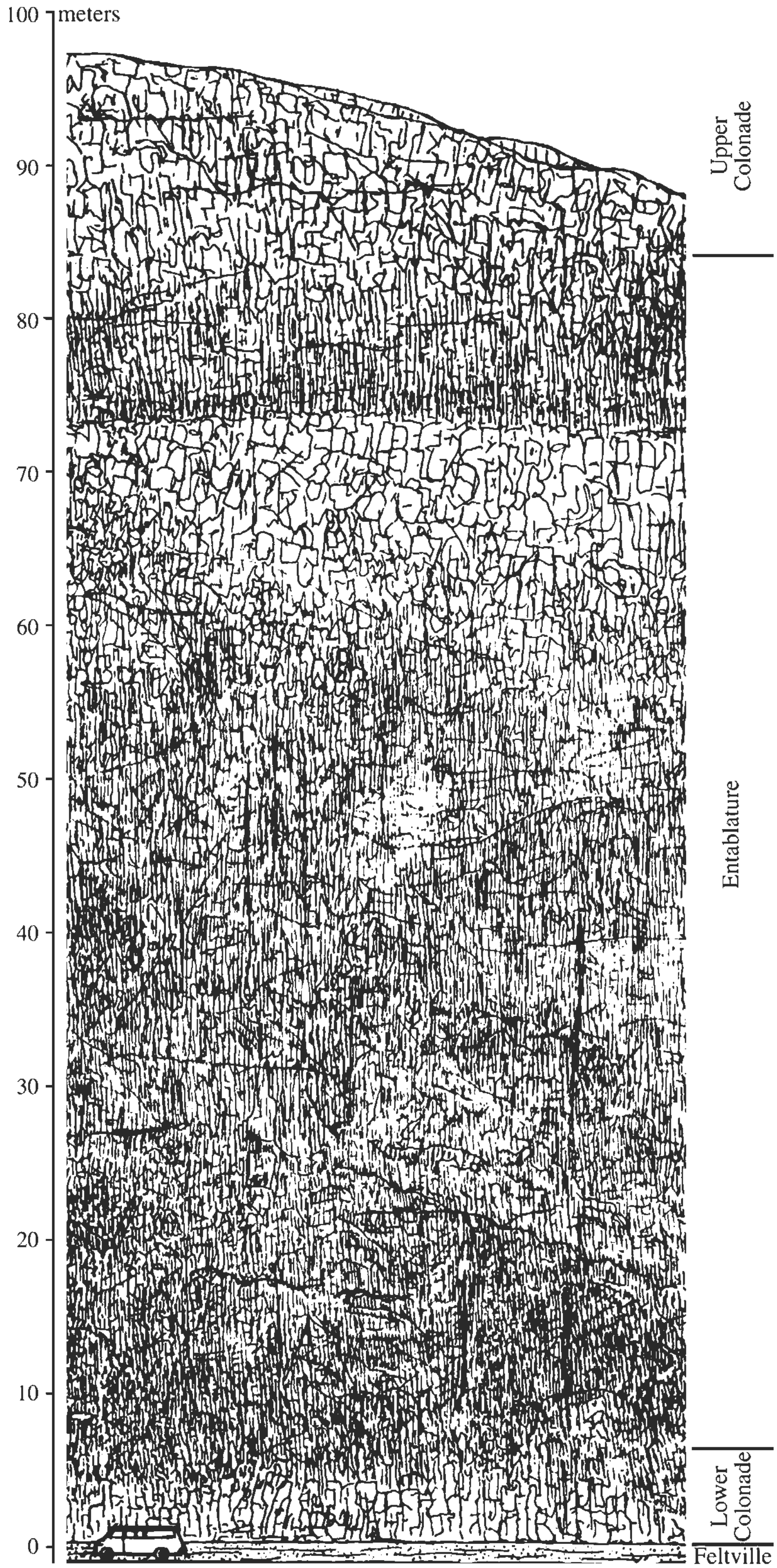


Figure 23. Composite drawing of the type section of the Preakness Basalt along Interstate Route 280 about 2.25 km west of the type section of the Orange Mountain basalt (Figure 22). Traced from a composite of a continuous series of photographs with the dip removed and compiled vertically (Olsen, 1980b).