Postapocalyptic greenhouse paleoclimate revealed by earliest Triassic paleosols in the Sydney Basin, Australia

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ABSTRACT

The Permian-Triassic boundary in the Sydney Basin of Australia is coincident with a pronounced decrease in $\delta^{13}C$ isotopic values of organic carbon, the last coals anywhere in the world for all of the Early Triassic (ca. 6 m.y.) time, and extirpation of the *Glossopteris* flora at the top of the Illawarra and Newcastle Coal Measures. Coal-bearing paleosols of the latest Permian represent extensive swamplands of the seasonally deciduous *Glossopteris* flora in a humid cold temperate lowland southwest of an Andean-style volcanic arc. Stone-rolls (large ribs of floor rock up into the coal) within some of the uppermost coals of the Permian coal measures can be interpreted as string bogs of a kind now found in cold climates at latitudes of 68°–70°, which is compatible with a paleomagnetically estimated paleolatitude of 65°–85°S for the Sydney Basin. Paleolatitude was not much different for earliest Triassic time, but paleosols of that age include Inceptisols and Entisols showing substantial chemical and textural weathering, more like soils now forming at latitudes of 40°–58° than those within polar circles. This anomalous high-latitude warmth set in at the Permian-Triassic boundary. Sedimentation rates increased at the boundary marked by geochemically unusual acidification of clay and a dramatic carbon isotopic excursion. These changes in environments and ecosystems can be explained by soil erosion following deforestation implied by the plant extinctions and abundant fungal remains at the Permian-Triassic boundary. Evidence from paleosols can now be added to that from paleontological and isotopic studies showing that disruption of the carbon cycle at the Permian-Triassic boundary resulted in a CO_2 or CH_4 post-apocalyptic greenhouse paleoclimate.

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INTRODUCTION

The greatest of all mass extinctions was at the Permian-Triassic boundary when life on this planet was almost lost and some 80%–97% of species became extinct (Stanley and Yang, 1994; Retallack, 1995). Long known as a major discontinuity in the fossil record of marine life (Erwin, 1993, 1994), this greatest of midlife crises has now been identified on land as well, due to advances in isotopic stratigraphy (Morante et al., 1994; Morante and Herbert, 1994; Morante, 1996), radiometric dating (Shaw et al., 1991; Retallack et al., 1993; Veevers et al., 1994a, 1994b; Conaghan et al., 1994), stratigraphy (Retallack et al., 1996b), paleobotany (Retallack, 1995; Poort, 1996; Wang, 1996), paleoentomology (Labandeira and Sepkoski, 1993), and vertebrate paleontology (Benton, 1987; Smith, 1995). Recent explanations for this extinction include oceanic anoxia (Wignall and Hallam, 1993; Wignall et al., 1995; Wignall and Twitchett, 1996; Isozaki, 1997), overturn of CO_2-rich deep ocean (Grotzinger and Knoll, 1995; Knoll et al., 1996), massive eruption of flood basalts of the Siberian Traps (Conaghan et al., 1994; Renne et al., 1995; Gurevitch et al., 1995), catastrophic release of methane from oceanic and permafrost clathrates (Erwin, 1993, 1994; Morante, 1996), and asteroid or comet impact (Hsü and McKenzie, 1990; Retallack et al., 1996a; Gorter, 1996).

This study aims to explore the nature of this mass extinction and associated environmental changes using primarily paleosols in the Sydney Basin, Australia. Paleosols are evidence of a concatenation of remarkable events at the Permian-Triassic boundary, including paleoclimatic warming, accelerated sedimentation rate, cessation of coal formation, leached boundary beds, and profound plant and animal extinctions. The warming at high latitudes, which I here term a postapocalyptic greenhouse, is indicated not only by paleosols studied here, but by fossil plants (Retallack, 1995, 1997a) and by isotopic studies of carbon and oxygen across the boundary (Holser and Magaritz, 1987; Magaritz and Holser, 1991; Holser et al., 1991; Morante, 1996).

MATERIALS AND METHODS

This research on paleoenvironmental change across the Permian-Triassic boundary in the Sydney Basin is aimed at characterizing variation throughout the basin (Figs. 1–4). Paleosol profiles were classified into pedotypes (of Retallack, 1994) in the field (Figs. 5–9; Table 1), with observations on Munsell color and reaction with dilute acid (following Retallack, 1990, 1997b). Thin sections of representative profiles were counted for 500 points using a Swift automatic point counter to determine the distribution of sand, silt, and clay, and of constituent minerals, with an accuracy of about 2% (Murphy, 1983). Bulk density was determined by the clod method using paraffin. Chemical analyses were from X-ray (XRF) fluorescence by Carol Lawson of Macquarie University, New South Wales (samples R1644-1680) and from inductively coupled plasma-fusion (ICP) spectroscopy by Bondar Clegg Inc., Vancouver, British Columbia (other samples), with ferric iron from ferrous ammonium sulfate titration and loss on ignition from 4 hr at 600 °C. Errors were estimated from multiple analyses of GSP-1 and AHV-1 for XRF data, of CANMET SY-3 and CANMET SO-2 for ICP, and from 10 replicates of specimen R1654 for bulk density. (Data and detailed paleosol descriptions are in the GSA Data Repository[1]).

My search for the Permian-Triassic boundary in nonmarine sections of the Sydney Basin was guided by the dramatic decrease in $\delta^{13}C$ of organic matter and of carbonate in numerous boundary sections worldwide (Magaritz et al., 1992; Wang et al., 1994; Yang et al., 1995; Yin, 1996). In the Sydney Basin, and other nonmarine basins across the

[1]GSA Data Repository item 9902, chemical and petrographic data, is available on request from Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301. E-mail: editing@geosociety.org. Web: http://www.geosociety.org/pubs/ftpyrs.htm.

Data Repository item 9902 contains additional material related to this article.

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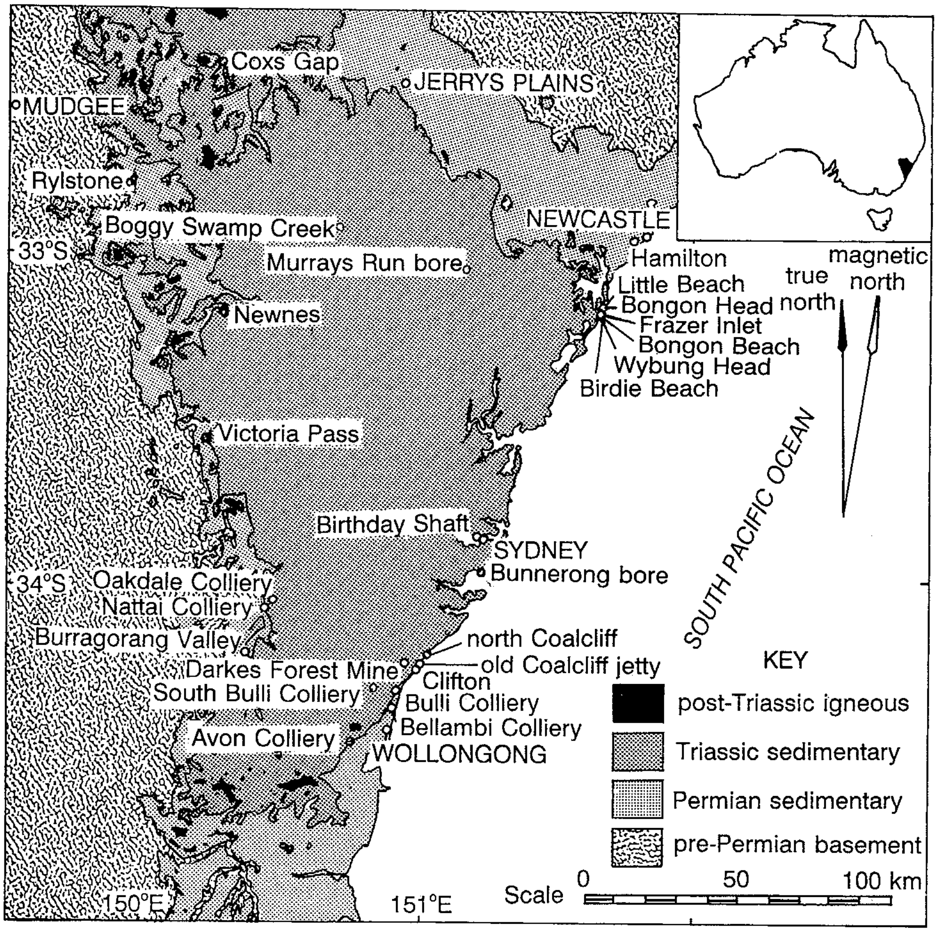


Figure 1. Mentioned localities and simplified geologic map of the Sydney basin, Australia.

continent to Western Australia, this chemostratigraphic marker horizon is at the base of the *Protohaploxypinus microcorpus* palynozone (Helby et al., 1987; Morante et al., 1994; Morante, 1996). This palynozone and isotopic shift also can be found in marine sequences of Western Australia (Morante, 1996) and within the upper sandstones of the Chhidhru Formation in the Salt Range of Pakistan (Balme, 1969; Foster, 1982; Baud et al., 1989), which in turn has been dated as Early Triassic using conodonts (Sweet, 1992; for an alternative correlation see Wignall and Hallam, 1993). Other palynological markers of the Permian-Triassic boundary include transient abundances of acritarchs, fern spores, and fungal remains (Eshet et al., 1995; Visscher et al., 1996), which also are found in the basal *P. microcorpus* palynozone in the Sydney Basin (Grebe, 1970; Retallack, 1995). The *P. microcorpus* palynozone corresponds with a zone of megafossil plants dominated by *Dicroidium callipteroides* (Retallack, 1980, 1995). Finally, this placement of the boundary is compatible with recent radiometric dating of the Permian-Triassic boundary in Siberia and China at 250 ± 0.3 Ma (Claoué-Long et al., 1991; Campbell et al., 1992; Renne et al., 1995), which is the interpolated age of the *P. microcorpus* zone from radiometrically dated Permian and Triassic volcanic rocks of New South Wales (Shaw et al., 1991; Retallack et al., 1993; Conaghan et al., 1994; Veevers et al., 1994a, 1994b).

Stratigraphically this places the Permian-Triassic boundary between the last coal of the Illawarra and Newcastle Coal Measures, and the base of the overlying Narrabeen Group. These rock units are subdivided differently in different parts of the basin. In southern coastal exposures of the boundary near the old Coalcliff jetty (Fig. 1), the Permian-Triassic boundary is at the top of the Bulli Coal and the base of the Coal Cliff Sandstone (Fig. 2, A and B: Harper, 1915; Hanlon, 1953). In northern coastal exposures at Wybung Head in Frazer State Park (Fig. 1) the boundary is at the top of the Vales Point Coal and the base of the Dooralong Shale (Fig. 2, C and D, Diessel, 1992). At both localities the basal thin (10–15 cm) bed of the Triassic indicated by the *Protohaploxypinus microcorpus* palynozone (Grebe, 1970; Helby, 1973) is a black, kaolinitic, claystone breccia (Figs. 2, 3, 6, 8, and 9).

Other boundary sequences were examined to the northwest near Coxs Gap (Fig. 3C), in the north central Sydney Basin in the Murrays Run DDH1 bore (Fig. 4A) and in the central Sydney Basin in the Bunnerong PHKB1 bore (Fig. 4B). Coxs Gap is just north of the Mount Coricudgy Anticline, which arbitrarily divides the Sydney Basin to the south from the Gunnedah Basin to the north (Mayne et al., 1974). The uppermost Permian unit at Coxs Gap is the Katoomba Coal. The basal Triassic unit there is best called Widden Brook Conglomerate, which is a transitional unit between the Caley Formation of the Sydney Basin and the Digby Formation of the Gunnedah basin (Uren, 1974; Tadros, 1993; Yoo, 1993). Nomenclature of the southern coalfield was used for the Bunnerong bore, and that of the Wollombi Coal Measures and north coastal Narrabeen Group for the Murrays Run bore (Mayne et al., 1974; Uren, 1980). The position of the boundary is revealed by evidence of both the *Glossopteris* and *Dicroidium callipteroides* floras at Coxs Gap (Retallack, 1980), *Glossopteris* flora, *P. microcorpus* palynofloras, and a decrease in the carbon isotopic composition ($\delta^{13}C_{org}$) of organic matter in the Murrays Run bore (Morante et al., 1994; Morante and Herbert, 1994; Morante, 1996), and megafossil plants including Late Permian *Vertebraria australis* and earliest Triassic *Isoetes beestonii* (Retallack, 1997a) in the Bunnerong bore.

ALTERATION OF PALEOSOLS AFTER BURIAL

Changes to the paleosols after burial need to be acknowledged before assessing their paleoenvironmental significance. Paleosols of the upper Narrabeen Group show clear evidence of three changes commonly altering soils shortly after burial: overall depletion in organic matter, chemical reduction (gleization) of oxides during decomposition of organic matter, and reddening of hydroxides by dehydration (Retallack, 1991). These are evident from an organic carbon content that is unusually low for soils (Morante, 1996), pervasive gray-green mottles around root traces, and locally red hues (Figs. 3 and 7, Retallack, 1977a). Early burial gleization also included widespread precipitation of siderite nodules and microspherulites (Retallack, 1977a, 1997c; Bai, 1988), which were already lithified and jointed before cover by overlying beds in the Coal Cliff Sandstone (Cook and Johnson, 1970). These various burial alterations have changed the appearance of what were formerly yellowish-brown to gray soils to the red-green mottled paleosols of today.

The paleosols have also been altered by burial processes including illitization, cementation, compaction, and jointing. The maximum preserved sequence of Triassic sedimentary rocks

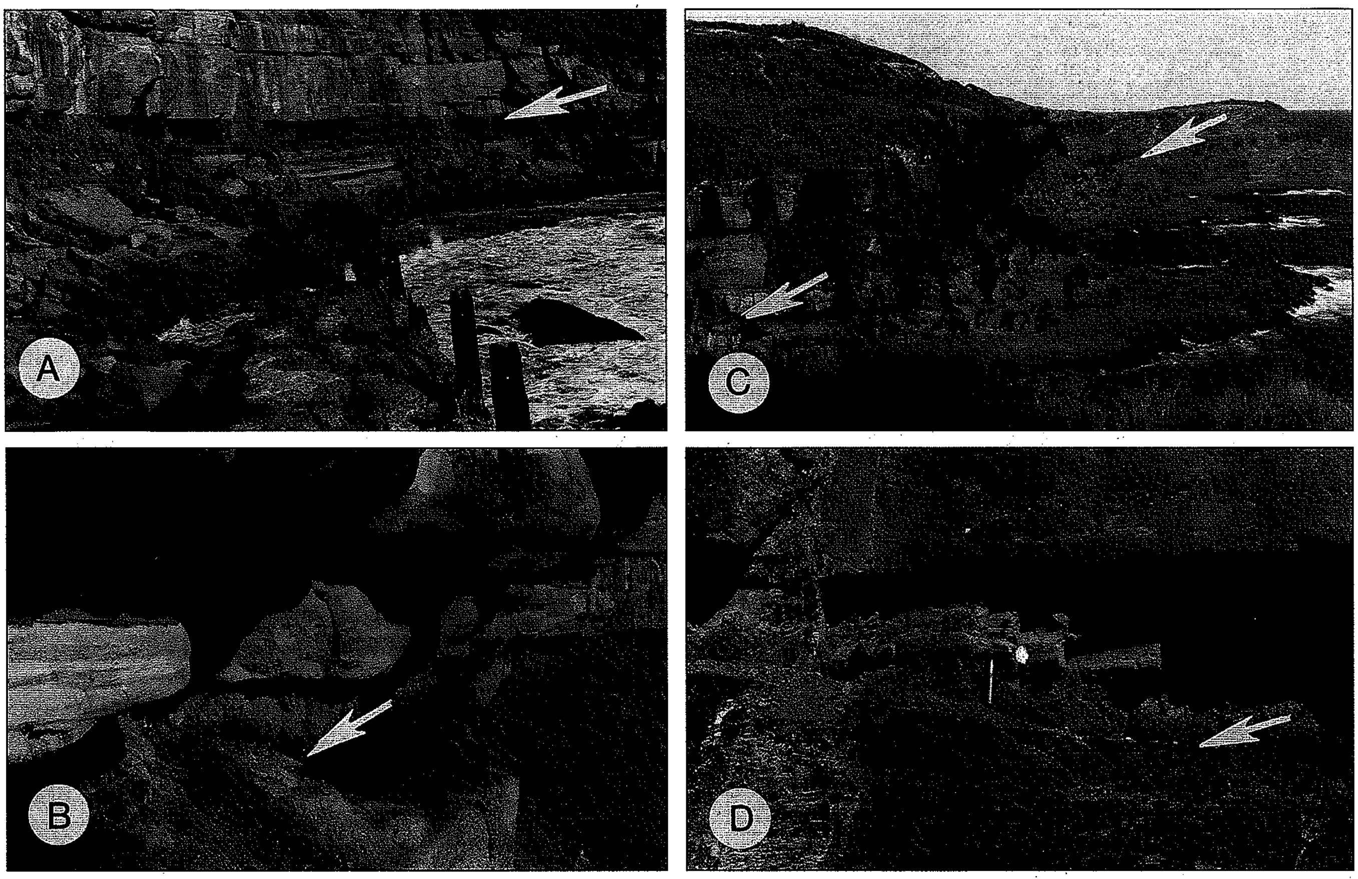


Figure 2. Permian-Triassic boundary (at arrow) in sea cliffs of the southern Sydney Basin at Coalcliff (A and B) and the northern Sydney Basin at Wybung Head (C and D). In both places the boundary is a contact between a thick coal seam and a kaolinitic claystone breccia at the base of a paleosol (B and D). Hammers give scale in close-up views.

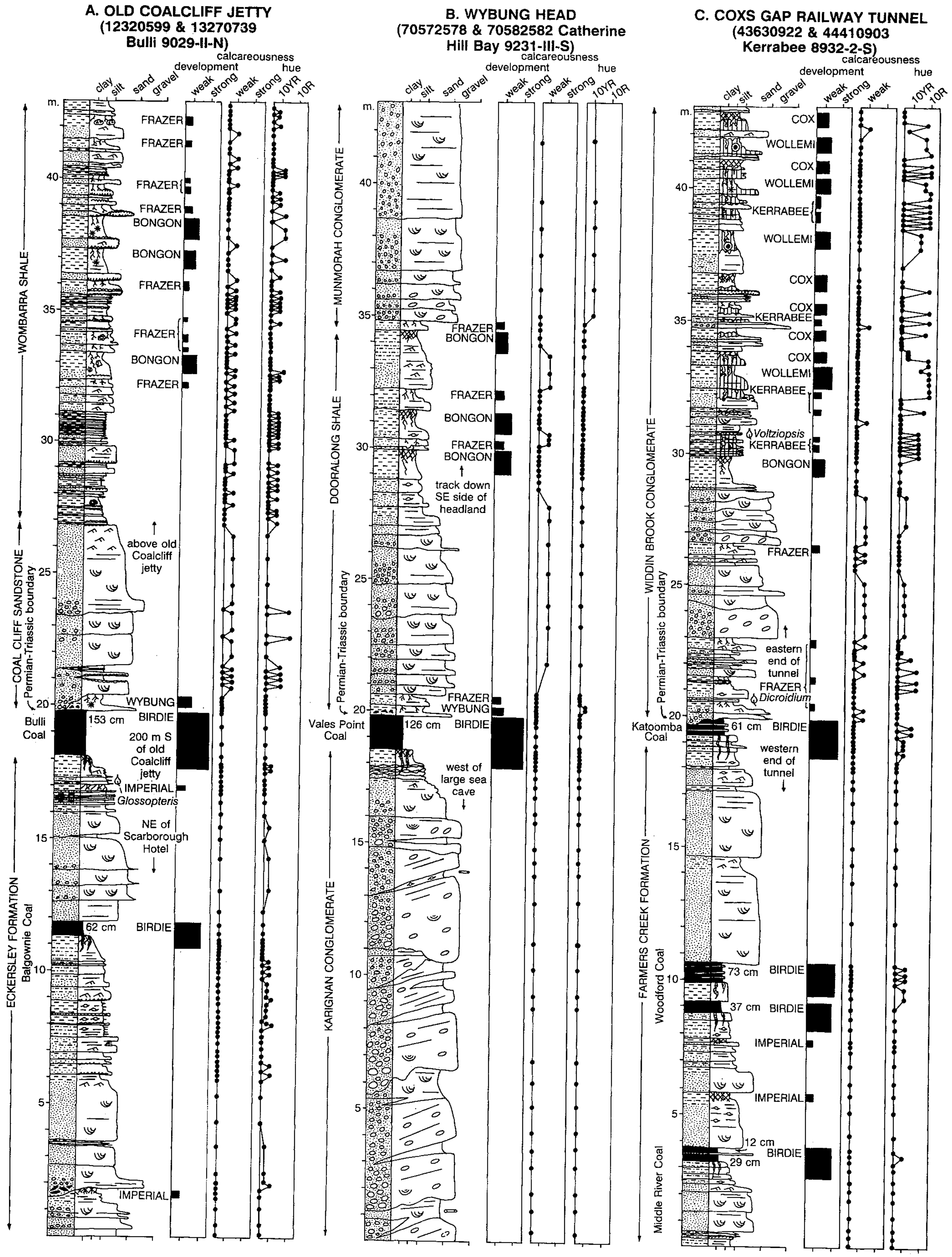
above the Bulli Coal is 1413 m (Uren, 1974; Stroud, 1974, Herbert, 1980). Late Triassic to early Jurassic shales some 800 m thick may have been eroded from above that. This missing section can be inferred from the shape of diatreme intrusions and the pollen content of their shale clasts (Crawford et al., 1980; Branagan, 1983), as well as from vitrinite reflectance and rank of latest Permian bituminous coals (Diessel, 1992; Faiz and Hutton, 1993). Both thermal modeling and oxygen isotopic studies of the lower Narrabeen Group indicate maximum burial depths of 1.5 km and maximum temperatures of 160–170 °C (Bai et al., 1990; Middleton, 1993). Compaction of the paleosols at this depth would have been negligible using the formula of Caudill et al. (1997). Expected under such conditions of burial are quartz overgrowth cements, seen especially in associated sandstones (Bai, 1988; Bai and Keene, 1996). Also found in sandstones is rare dawsonite cement (Baker et al., 1995). Illitization also would be expected at such depths (Eberl et al., 1990), and is quite far advanced, with sharp illite peaks on XRD traces and little mixed layer illite-smectite remaining (Loughnan, 1963, 1966; Bai, 1988; Retallack, 1977a). Illitization and concomitant potash enrichment is no more than 3.25 wt%, part of which is due to common detrital muscovite in these rocks. Illitization has, however, brightened the birefringence fabric of these paleosols. As in other cases (Retallack and Krinsley, 1993), this appears to have been a heightening of preexisting pedogenic birefringent streaks. These rocks are little deformed by folding and have low dips, but have been locally jointed and faulted (Conaghan, 1984; Agrali, 1987, 1990; Memarian, 1993).

PALEOCLIMATIC WARMING

Paleoclimate can be inferred from both specific features of the paleosols and from their comparison with Quaternary soils and soilscapes (Mack and James, 1992, 1994; Retallack, 1997b). Such studies indicate a profound paleoclimatic warming, but not much change in rainfall across the Permian-Triassic boundary in the Sydney Basin (Table 2).

Stone-rolls (Fig. 10), clastic dikes, and flame structures of the uppermost Permian coal measures (Cook and Johnson, 1970; Diessel, 1992, Figs, 5.17 and 6.23) have been interpreted as evidence for deformation in the active layer of permafrosted soils and thus a frigid, periglacial paleoclimate (Conaghan et al., 1994). Stone-rolls are large ribs of rock extending upward into the base of the Bulli and Wongawilli Coals (Conaghan et al., 1994), and perhaps also the Woonona Coal (Bowman, 1970). Stone-rolls within the Bulli Coal average 0.89 ± 0.33 m high, 3.8 ± 3.77 m wide, and 206 ± 217 m long (means and standard

Figure 3. Sequences of paleosols across the Permian-Triassic boundary in outcrops to the south, north, and west of the Sydney Basin respectively, at Coalcliff (A) near Clifton, Wybung Head (B) in Frazer State Park, and Coxs Gap railway tunnel (C) near Kerrabee. Lithological key as in Figure 5. Scales of grain size, calcareousness, and development are after Retallack (1990, 1997b), and hues are from a Munsell color chart.


A. OLD COALCLIFF JETTY
(12320599 & 13270739
Bulli 9029-II-N)
B. WYBUNG HEAD
(70572578 & 70582582 Catherine
Hill Bay 9231-III-S)
C. COXS GAP RAILWAY TUNNEL
(43630922 & 44410903
Kerrabee 8932-2-S)
calcareousness
development
hue
clay
silt
sand
gravel
weak
strong
10YR
10R
m.
WOMBARRA SHALE
COAL CLIFF SANDSTONE
Permian-Triassic boundary
Bulli
Coal
ECKERSLEY FORMATION
Balgownie Coal
FRAZER
BONGON
WYBUNG
BIRDIE
IMPERIAL
153 cm
62 cm
above old
Coalcliff
jetty
200 m S
of old
Coalcliff
jetty
Glossopteris
NE of
Scarborough
Hotel
MUNMORAH CONGLOMERATE
DOORALONG SHALE
KARIGNAN CONGLOMERATE
Vales Point
Coal
126 cm
track down
SE side of
headland
west of
large sea
cave
WIDDIN BROOK CONGLOMERATE
Katoomba
Coal
FARMERS CREEK FORMATION
Woodford Coal
Middle River Coal
COX
WOLLEMI
KERRABEE
Voltziopsis
Dicroidium
eastern
end of
tunnel
western
end of
tunnel
61 cm
73 cm
37 cm
12 cm
29 cm

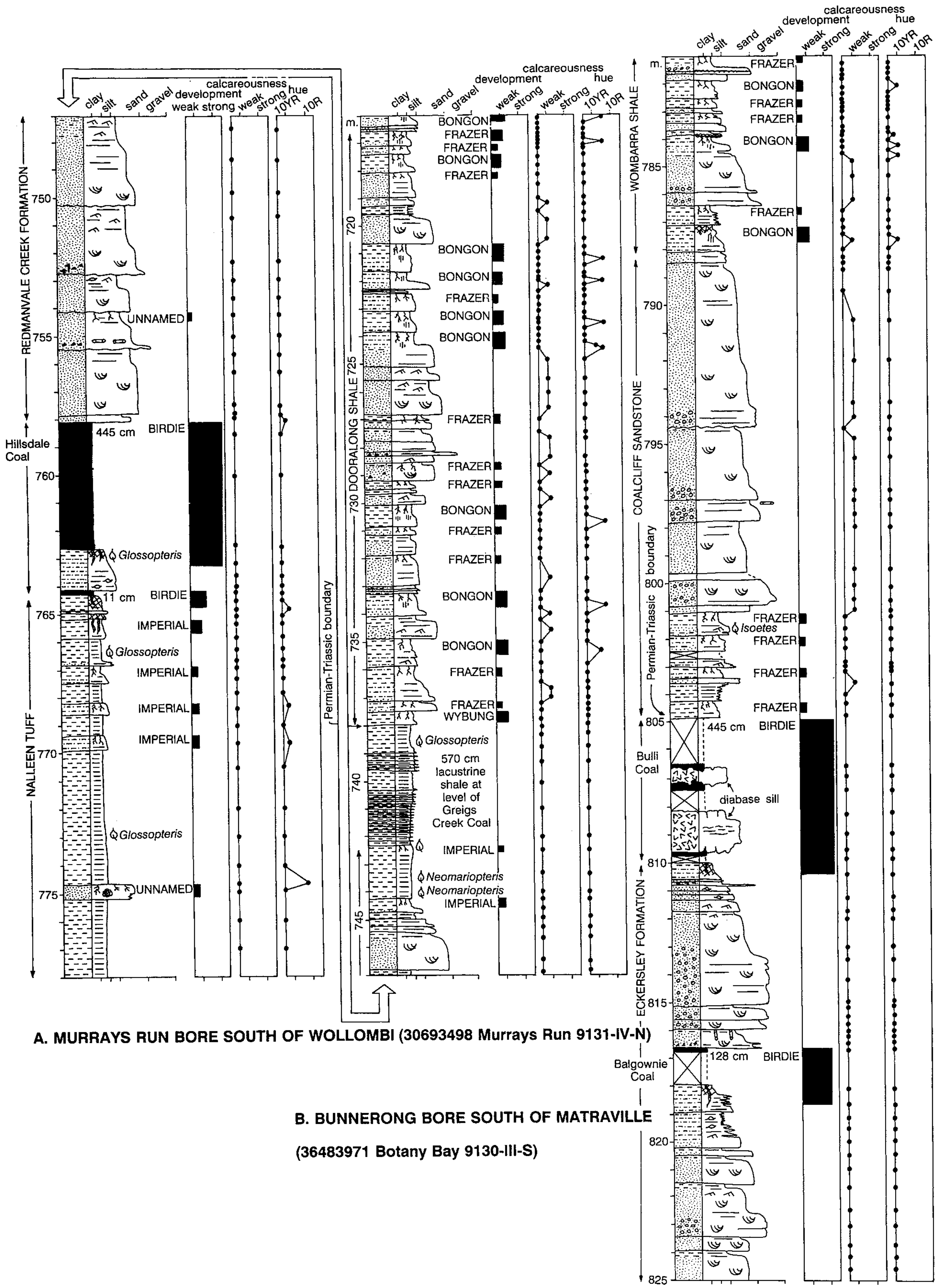


Figure 4. Sequences of paleosols across the Permian-Triassic boundary in deep bore holes, Murrays Run (A) near Wollombi, and Bunnerong (B) near Matraville. Lithological key and other conventions as for Figure 5.

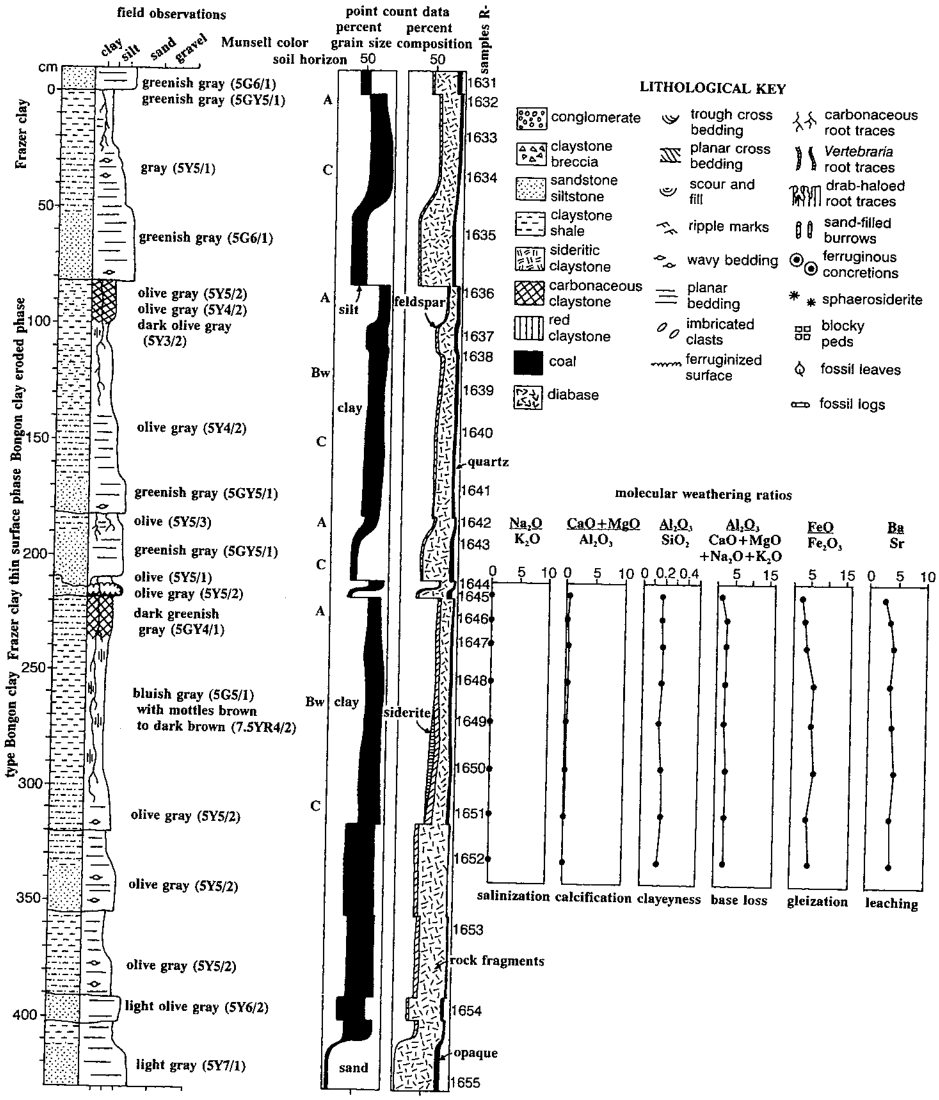


Figure 5. Measured section, Munsell colors, soil horizons, grain size, mineral composition, and molecular weathering ratios of the Early Triassic Bongon and Frazer paleosols of the Dooralong Shale at Wybung Head (grid reference 70582582, Catherine Hill Bay 9231-III-S). This is at 27–32 m in Figure 3B.

deviations of 1000 or more measurements by Agrali, 1990).

The stone-rolls were not formed by postdepositional deformation of the coal floor, but were originally present at the base of the peat, for the following five reasons. (1) Coal interfingers with claystone and shale of the stone-roll (Fig. 10A). (2) Sideritic claystone bands (Fig. 10B) and individual coal plies (seam bands) (Diessel and Moelle, 1970) onlap the stone-rolls, whereas both stone bands and coal plies in the upper part of the seam, as well as the roof shales and sandstone are little deformed. (3) Some stone-rolls have a different lithological composition from planar seat earths to the coal (Diessel and Moelle, 1970). (4) Stone-rolls are commonly cemented by sideritic cement, which can be shown to have formed before deposition of the roof by sideritic clasts within the overlying fluvial sandstone (Cook and Johnson, 1970). (5) Stone-rolls preserve little-deformed, large, chambered roots of *Vertebraria*, despite randomly arranged slickensided clay skins, here interpreted as pedogenic rather than tectonic.

Two explanations have been advanced for the stone-rolls compatible with this evidence for origin during early stages of peat accumulation: (1) deposits of creeks (Diessel and Moelle, 1970) or (2) vaulting by permafrost (Conaghan, 1984; Conaghan et al., 1994). Neither of these explanations withstands close scrutiny. Maps of the stone-rolls and their oblique tie-rolls (rock ridges connecting parallel stone-rolls) show striking linearity, and lack of variance in orientation consistently perpendicular to stream paleocurrents (Agrali, 1987, 1990), unlike river or creek deposits, especially "senile drainage" envisioned by Diessel and Moelle (1970). Slabs and thin sections prepared from oriented specimens of the rolls for this study (Fig. 10) failed to reveal frost brecciation like that found in modern permafrost soils (van Vliet Lanoë, 1985) or string bogs (Seppälä, 1988). Nor were there clastic dikes like those in coals interpreted as permafrost string bogs of mid-Permian age in southern Victoria Land, Antarctica (Krull and Retallack, 1995).

There is no support for permafrost climate from paleobotany or coal petrography. The Bulli and Wongawill stone-rolls contain little-deformed woody root traces of *Vertebraria* (Fig. 10) and the coals are rich in wood (vitrinite: Smyth, 1970; Shibaoka and Smyth, 1975; Cook, 1975), resin from wood (Diessel, 1992, Fig. 2.18) and glossopterid leaves (Diessel, 1992, Fig. 2.12,13), whereas permafrost string bogs envisaged by Conaghan et al. (1994) are developed in domed (ombrotrophic) peats of *Sphagnum* moss and other herbaceous plants (palsamires of Ruuhijärvi, 1983). Fossil mosses are known from Permian coal measures of Siberia (Neuburg, 1960), Antarctica (Smoot and Taylor, 1986), and South Africa (Anderson and Anderson, 1985), but not yet from the Sydney Basin, where lycopods (*Selaginella*) and horsetails (*Phyllotheca*) are the principal Late Permian herbaceous plants (Townrow, 1956, 1968). Glossopterids that dominate the latest Permian coal measures were deciduous trees (David, 1907; Retallack, 1980; Diessel, 1992). Permineralized trunks (*Dadoxylon* and *Araucarioxylon*) are commonly 10–20 cm in diameter, and to 50 cm in diameter within coals lacking stone-rolls of the upper Newcastle Coal Measures (Diessel, 1992). Four fossil logs within the A horizon of the Bulli Coal at Coal Cliff had compacted widths of 8.8, 18.0, 20.5, and 23.5 cm. Compressed width is equal to the former diameter following Walton's (1936) compaction hypothesis. The height of the stoutest trees from the Bulli Coal can be estimated as about 16 m using regressions of height versus stem diameter derived from a variety of cool-temperate living trees (Whittaker and Woodwell, 1968). The strong growth rings of this fossil wood and the local abundance of leaves within the spring layer of varved shales has been used to infer that these plants were seasonally deciduous (Retallack, 1980). Despite recent doubts about this (Pigg and Trivett, 1994), further evidence for seasonal leaf-fall comes from callused abscission scars at the base of fossil leaves (Anderson and Anderson, 1985). In polished thick sections of coals, glossopterid leaves also show alteration of

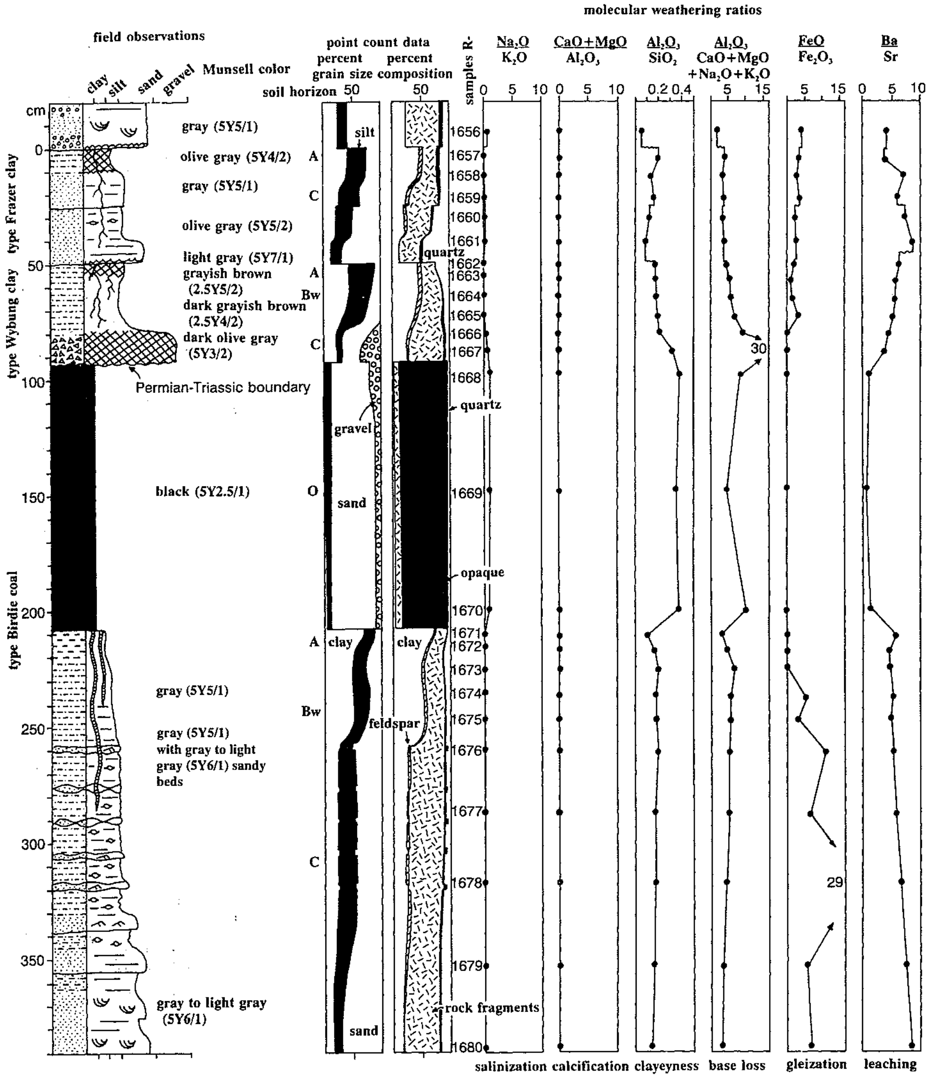


Figure 6. Measured section, Munsell colors, soil horizons, grain size, mineral composition, and molecular weathering ratios of the Permian–Triassic Frazer, Wybung, and Birdie paleosols of the Dooralong Shale above the Vales Point Coal of the Karignan Conglomerate at Wybung Head (grid reference 70572578 on Catherine Hill Bay 9231-III-S). This is at 16–21 m in Figure 3B. Lithological key as for Figure 5.

mesophyll to semifusinite, interpreted by Diessel (1992) to be due to freeze-drying in autumnal banks on the ground during winter before incorporation in peat during the following summer.

The modest mineral content and reflected-light petrography of the coals are also evidence against domed peats, indicating instead that ground water was more important than rain water in these swamps (rheotrophic of Moore and Bellamy, 1974). For example, coals of the upper Newcastle Coal Measures show an unusual combination of medium to low tissue preservation index and very low gelification index, which Diessel (1992) interpreted as evidence of seasonal freezing of wood and leaves. The Bulli and Wongawill Coals lack the high inertinite and low vitrinite content of Late Triassic coals in Tasmania interpreted as former permafrost string bogs (palsamires, Smyth, 1980).

Despite these problems, there may be elements of truth in both creek and permafrost explanations. The stone-rolls are here envisaged as seasonal ponds (mud-bottom flarks or lateral depressions) between ridges of string bogs. Nonpermafrost string bogs can form on very slight slopes through accumulation against obstacles of plant debris transported by slow laminar flow of water or blown across seasonally icy ponds or snow banks (Fig. 11). This explanation accounts for the modest internal deformation of the stone-rolls and the onlapping coal plies and stone bands (Fig. 10). By this explanation, the domed form is in part due to much greater compaction of peat than clay (Fig. 11). Thus, I envisage the Bulli and Wongawilli stone-rolls to have formed string bogs similar to the Finnish Pohjanmaa type of aapamire (string bog) (Ruuhijärvi, 1983). This kind of patterned bog can be eutrophic and support broadleaf deciduous trees, such as birch (*Betula michauxi*) as tall as 10 m. Aapamires form well south of the polygonal-patterned peats of the Arctic Circle and south also of palsamires (permafrost-cored peat domes) of the zone of discontinuous permafrost. Aapamires are now found at latitudes of 63°–67°N in Finland (Ruuhijärvi, 1983) and northwestern Russia (Botch and Masing, 1983), but are at lower latitudes of 50°–54°N in continental regions with less maritime influence, such as northern Canada (Zoltai and Pollett, 1983). Generally similar peaty soilscapes are now found at latitudes of 65°–72° north on the coast of the White Sea and Pechora basin of northern Russia (map unit Ox7-a of Food and Agriculture Organization, 1981) and at latitudes of 51°–59° north in Alberta, Labrador, Ontario, and Quebec, Canada (map units Ox1-a, Ox6-a of Food and Agriculture Organization, 1975). The higher of these latitudes may be appropriate for the latest Permian of New South Wales, because of marine influence in the Erins Vale Formation lower in the Illawarra Coal Measures (Mayne et al., 1974), likely eustatic control of coal measure sedimentation (Herbert, 1997a), and the likelihood of warm southerly ocean currents along the Late Permian coast (Crowley, 1994).

Such high paleolatitudes are compatible with those determined from paleomagnetism, although specific estimates for Sydney vary from 85°S with the pole in central New South Wales (Embleton, 1984), to 65°S with the pole in or near palinspastically restored terranes of New Zealand off the coast of Antarctica (Scotese and Denham, 1988; A. G. Smith for Barrett, 1991). From this it can be inferred that latest Permian climate in the Sydney Basin was cold temperate and only marginally frigid. This is colder than inferred by some paleoclimatic models (Loughnan, 1991; Kutzbach and Ziegler, 1993) but not others (Fawcett et al., 1994).

Climate was very different in earliest Triassic time. Paleosols of the lowest Narrabeen Group show common clay skins (Fig. 12A), microspherules of siderite (sphearosiderite; Fig. 12, C and D), and opaque mottles, indicating an intensity of soil formation not seen in latest Permian paleosols, and quite unlike the little-weathered, clay-poor soils of frigid regions (Tedrow, 1977; Bockheim and Ugolini, 1990). Climate remained within the bounds of forest biomes, because fossil logs as wide as 20 cm are found in the lower Narrabeen Group in Oakdale and Bulli Collieries and on the coast near Burning Palms. Using Walton's (1936) rule to convert compressed width to diameter and linear regressions of diameter to height of trees (Whittaker and Woodwell, 1968)

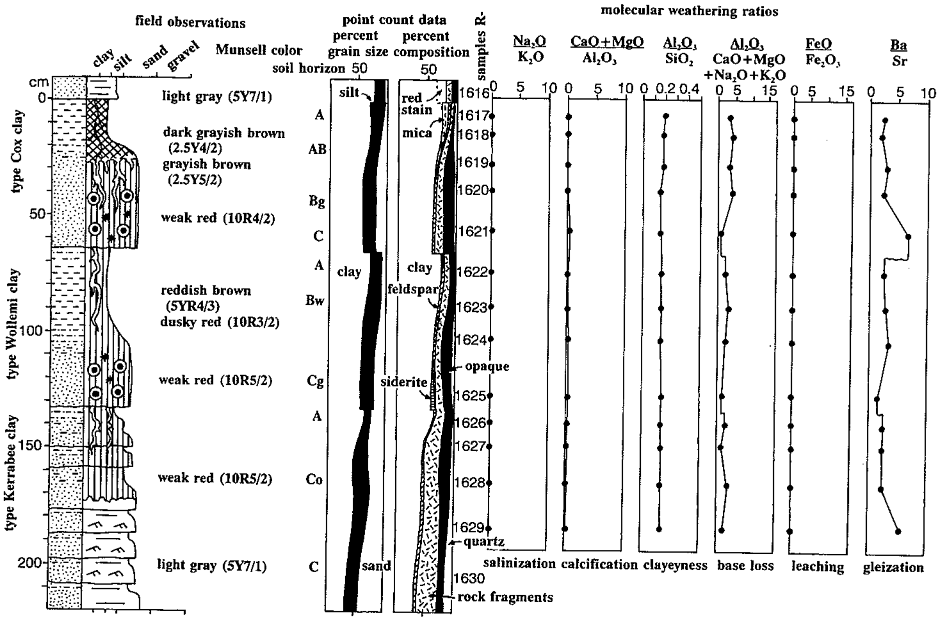


Figure 7. Measured section, Munsell colors, soil horizons, grain size, mineral composition, and molecular weathering ratios of the Early Triassic Kerrabee, Wollemi, and Cox paleosols of the Widdin Brook Conglomerate at the eastern end of Coxs Gap railway tunnel (grid reference 44410903, "Kerrabee" 8933-I-S). These are at 32–34 m in Figure 3C. Lithological key as for Figure 5.

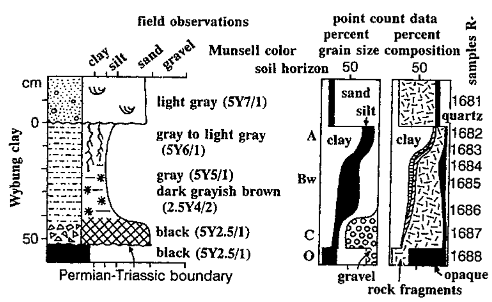


Figure 8. Measured section, Munsell colors, soil horizons, grain size, and mineral composition of the earliest Triassic Wybung paleosol of the Coalcliff Sandstone at Coalcliff (grid reference 13270739, Bulli 9029-II-N). This is at 19–20 m in Figure 3A. Lithological key as for Figure 5.

gives an estimate of former tree height of about 15 m. Climate probably remained highly seasonal, as indicated by silt-clay banded clay skins (Fig. 12A) and ferruginized concretions and microspherulitic siderite (Fig. 12, C and D). Seasonal climate for the lower Narrabeen Group is also indicated by strong growth rings in fossil wood (Baker, 1931; Burges, 1935). The harsh season of arrested growth was probably snowy, rather than dry, given the lack of calcareous nodules and overall deep weathering of these paleosols (high alumina/silica, alumina/bases, and Ba/Sr ratios), which is impressive because relict bedding indicates times for formation of no more than a few thousand years for these paleosols (Retallack, 1990).

In the soil classification of Food and Agriculture Organization (1974), paleosols of the lower Narrabeen Group are mainly Cambisols (Bd, Bg) and Gleysols (Gd), comparable to soils at middle latitudes (35°–58°N) in the South Island of New Zealand (map unit Bd3-2a, Bd3-2bc, Bd3-2c of Food and Agriculture Organization 1978b), in Britain, France, Austria, Belgium, and Switzerland (Bd42–1/2bc of Food and Agriculture Organization, 1981), in the Bureya Basin of far eastern Russia (Bd3–2ab of Food and Agriculture Organization, 1978a), in coastal Korea (Bd3–2b of Food and Agriculture Organization, 1978a), and in North America from the east ("sunshine coast") of Vancouver Island, to southern Quebec, West Virginia, New York, Pennsylvania, and New Jersey (Bd3-2a, Bd3-2abc, Bd20-2a, Bd20-2ab, Bd20-2b of Food and Agriculture Organization, 1975). These are regions of humid, mixed conifer-hardwood forests and cold snowy winters. In terms of climatic life zones of Walter (1985), the Permian-Triassic boundary appears to have been a significant step from cold-temperate and marginally frigid swampland to cool-temperate broadleaf and conifer forest.

Cool temperate broadleaf and conifer forests are anomalously warm for latitudes of 65°–85°S determined paleomagnetically (Embleton, 1984; Scotese and Denham, 1988; A. G. Smith for Barrett, 1991), and so indicate not only a warmer Early Triassic climate, but an equator-pole temperature gradient different from that of today. The latitude of Victoria Land and the Alaskan North Slope today are 65°–85°, where soils are thin, silty, and little weathered (Tedrow, 1977; Bockheim and Ugolini, 1990). Anomalous polar warmth is a notable feature of some models of Mesozoic paleoclimates (Ziegler et al., 1993), although other models give anomalously cool Early Triassic polar regions (Wilson et al., 1994). From evidence presented here, anomalous polar warmth dates back to earliest Triassic time.

Humid climate has been inferred for Late Permian time from the accumulation of thick coals (Retallack, 1980), and for Early Triassic time from the noncalcareous and deeply weathered nature of paleosols in Queensland (Jensen, 1975), New South Wales (Retallack, 1977b, 1997c), and Antarctica (Barrett and Fitzgerald, 1986). Indications of a rainy climate for the Sydney Basin come from the low content of alkalies and alkaline earths, both in absolute value and compared with parent materials of the paleosols. Coal-bearing paleosols would have been somewhat insulated from regional climate by waterlogged peat, but the deeply penetrating *Vertebraria* roots in their underclay are evidence for a time before peat accumulation when the underclay was freely drained and influenced by rainfall. Thus, geochemical trends in the underclays also have paleoclimatic significance. Enrichment of barium with respect to strontium and alumina compared with alkalies, alkaline earths, and silica also indicate a humid climate by comparison with soils today (Chesworth, 1992; Retallack, 1997c). Both Permian underclays and Triassic paleosols, with the exception of the profile immediately above the Permian-Triassic boundary, show values of alumina/bases that are higher in their B horizons than their C horizons. Quantitative estimates of former rainfall can be gained from the molecular ratio bases/alumina (B) of the Bt horizons of North American soils, which can be shown to be related to mean annual rainfall (P, mm) according to the following formula (Ready and Retallack, 1995):

$$P = -759B + 1300, \qquad (1)$$

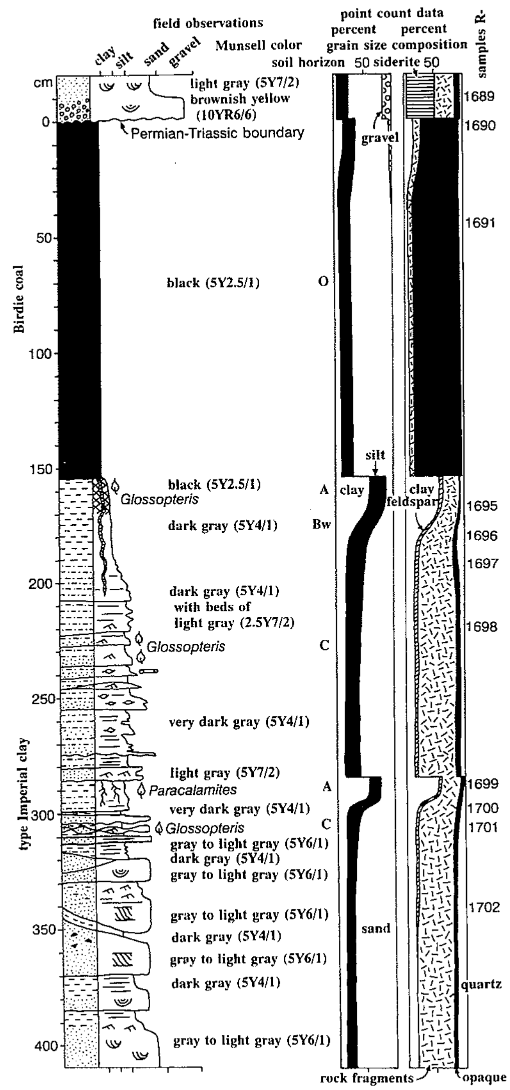


Figure 9. Measured section, Munsell colors, soil horizons, grain size, and mineral composition of the latest Permian Birdie and Imperial paleosols of the Eckersley Formation at Coalcliff (grid reference 13250722, Bulli 9020-II-N). This is at 16–19 m in Figure 3A. Lithological key as for Figure 5.

with correlation coefficient (r) of 0.7 and standard deviation (v) of ±174 mm. This equation applied to two samples of a latest Permian coal-bearing paleosol (Birdie pedotype) indicates rainfall of 1188 and 1168 ± 174 mm, which is similar to that calculated for the earliest Triassic Wybung clay paleosol of 1189 and 1210 ± 174 mm. Similar calculations based on the chemical composition of Bongon and Wollemi paleosols indicate rainfall of 963–1012 ± 174 mm and 987–1047 ± 174 mm, respectively. In round figures, this amounts to precipitation of 1000–1300 mm for both latest Permian and earliest Triassic time, and 800–1200 mm for slightly younger (ca. 100 ka.) Early Triassic time. Thus, climate remained broadly humid across the Permian-Triassic boundary, and a shift to arid climates inferred for Europe (Mader, 1990), China (Wang, 1993,) and South Africa (Smith, 1995) was not also found in the Sydney Basin.

DEEPLY LEACHED BOUNDARY BRECCIAS

The first few meters of the earliest Triassic Narrabeen Group are characterized by unusual abundances of minerals resistant to weathering. Quartz is unusually abundant in paleochannels and weakly developed paleosols immediately above the Permian-Triassic boundary (Figs. 6 and 8). In cliffs high above Little Beach north of Wybung Head (Fig. 1) a distinctive white, quartz-rich sandstone overlies the Vales Point Coal (David, 1907). Kaolinite is also locally abundant in the base of the Wybung paleosol, which includes a distinctive basal claystone breccia at both Coalcliff and Wybung Head (Fig. 12B). The kaolinite clasts have microstructures characteristic of soils, including omnisepic plasmic fabric and illuviation argillans of Brewer (1976). The degree of leaching of bases from the claystone breccias is quite extraordinary (Fig. 6). A value of 30 for the molar ratio of alumina/bases would be unusual even for Oxisols, which are deeply weathered tropical soils formed over long periods of time (100 k.y. or more, Retallack, 1997b). Despite the anomalous deeply weathered sediments at this level in the Sydney Basin, there is no comparably profound weathering of either overlying or underlying paleosols. The claystone breccias record an unusual pulse of soil erosion at the Permian-Triassic boundary.

ACCELERATED SEDIMENTATION RATE

Estimates of sediment accumulation rates for paleosols are based on the assumption that these sequences represent long periods (hundreds to thousands of years) of soil formation punctuated by short periods (days and weeks) of sedimentation (Retallack, 1984; Bown and Kraus, 1993). The time represented by each paleosol can be calculated by comparison with radiocarbon-dated estimates for the time it took to form soils of comparable development on the present landscape. Soils comparable to Early Triassic paleosols of the Sydney Basin are found in humid forested regions of Pennsylvania, United States, and in the South Island of New Zealand (F.A.O., 1975, 1978b), where Entisols form over periods of less than 40 yr, Inceptisols form over 0.6–1.6 k.y. and Alfisols and Spodosols form over 3–22 k.y. (Ciolkosc et al., 1990; Tonkin and Basher, 1990). A terrace soil along Cave Stream in an area of New Zealand receiving 1470 mm mean annual rainfall has a 6% subsurface enrichment in clay in a horizon (Bt) about 30 cm thick within less than 7980 ± 80 radiocarbon years (Tonkin and Basher, 1990). This is a degree of development in excess of that seen in the Bongon pedotype, which is the thickest and most clay-enriched of the Early Triassic paleosols. About 5 k.y. would be the maximum likely time to form a Bongon profile, and proportionally less can be assigned to each of the other pedotypes using such features as preservation of relict bedding, and degree of chemical and mineral weathering as indices of development (Table 2).

TABLE 1. LATEST PERMIAN AND EARLIEST TRIASSIC PALEOSOLS OF THE SYDNEY BASIN

Pedotype	Age	Diagnosis	Australian classification (Stace et al., 1968)	Australian system (Isbell, 1993)	U.S. taxonomy (Soil Survey Staff, 1997)	F.A.O.* World Map (F.A.O., 1974)
Birdie	Late Permian	Thick coal over gray siltstone with *Vertebraria*	Acid peat	Fibric organosol	Cryofibrist	Gelic histosol
Bongon	Early Triassic	Gray claystone over brown-mottled sideritic silty claystone	Humic gley	Redoxic hydrosol	Humaquept	Humic gleysol
Cox	Early Triassic	Carbonaceous claystone over red bedded siltstone	Wiesenboden	Leptic tenosol	Haplaquept	Gleyic cambisol
Frazer	Early Triassic	Gray claystone with relict bedding and roots	Alluvial soil	Oxyaquic hydrosol	Hydraquent	Dystric fluvisol
Imperial	Late Permian	Dark gray shale with roots	Alluvial soil	Redoxic hydrosol	Cryaquent	Dystric fluvisol
Kerrabee	Early Triassic	Red-green mottled siltstone with relict bedding	Brown clay	Stratic rudosol	Udifluvent	Dystric fluvisol
Wollemi	Early Triassic	Green mottled silty clay over red claystone	Brown earth	Orthic tenosol	Dystrochrept	Dystric cambisol
Wybung	Earliest Triassic	Gray claystone over coal breccia	Gray clay	Redoxic hydrosol	Humaquept	Dystric gleysol

*F.A.O.—Food and Agriculture Organization of UNESCO (United Nations Educational, Scientific, and Cultural Organization).

TABLE 2. PALEOENVIROMENTAL INTERPRETATION OF LATEST PERMIAN AND EARLIEST TRIASSIC PALEOSOLS OF THE SYDNEY BASIN

Pedotype	Paleoclimate	Former vegetation	Paleotopography	Parent material	Time to form
		Latest Permian paleosols			
Birdie	Humid (1000–1300 mm/yr^{-1}), cold-temperate, seasonally snowy	Peat swamp forest, including *Glossopteris kingii* with attached fertile structure of *Senotheca kingii* (dominant), *G. gladiforma* and associated fertile *Squamella australis*, *Glossopteris chevronata* attached to *Austroglossa walkomii*, and *Vertebraria australis*	Permanently inundated swamp of extensive coastal plain	Illitic shale, quartzo-feldspathic siltstone	23–85 k.y.
Imperial	Not sufficiently developed to indicate paleoclimate	Early successional marsh, including *Paracalamites australis* (dominant) and *Neomariopteris lobifolia*	Point-bar swales and lake margins	As above	1–100 yr
		Earliest Triassic paleosols			
Bongon	Humid (800–1200 mm/yr^{-1}), cool-temperate	Wet mid-successional forest	Clayey, seasonally dry waterlogged flood plain	Illitic shale, quartzo-feldspathic siltstone	2–5 k.y.
Cox	Humid, seasonally cold	Lowland, periodically inundated forest	Clayey, seasonally inundated flood plain	As above	1–2 k.y.
Frazer	Not sufficiently developed to indicate paleoclimate	Early successional woody scrub including *Dicroidium callipteroides* (dominant), *Paracalamites* sp., *Cladophlebis carnei* and *Merianopteris* sp.	River point bars within lowland flood plain	As above	10–100 yr
Kerrabee	Not sufficiently developed to indicate climate	Early successional woodland, with *Voltziopsis wolganensis* (dominant), *Isoetes beestonii*, and *Cladophlebis* sp.	Well-drained river point bars	As above	1–400 yr
Wollemi	Humid (800–1200 mm/yr^{-1}), seasonally cold	Lowland forest	Well-drained clayey flood plain	As above	1–2 k.y.
Wybung	Humid (1000–1300 mm/yr^{-1}), cool-temperate	Wet mid-successional forest	Clayey flood plain	Kaolinitic claystone breccia and coal	2–3 k.y.

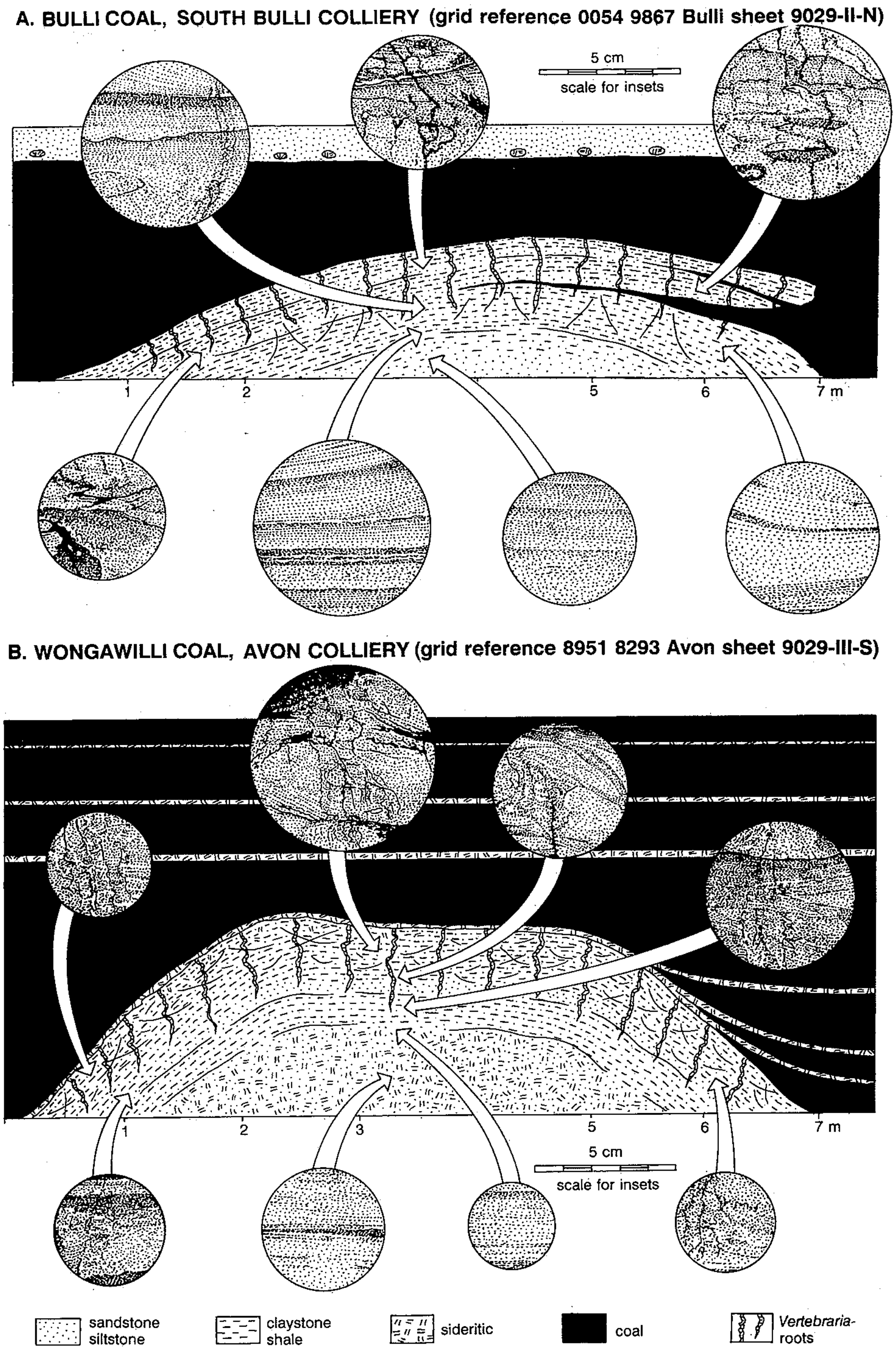


Figure 10. Field sketch and sedimentary structures visible in oriented sawn slabs of Late Permian stone-rolls from underground coal mines in the southern Sydney Basin, showing clear *vertebaria* roots but no frost brecciation. (A) Bulli Coal, the uppermost seam of the Illawarra Coal Measures. (B) Wongawilli Coal, the fifth coal from the top of the Illawarra Coal Measures. Cross seams are drawn with no vertical exaggeration.

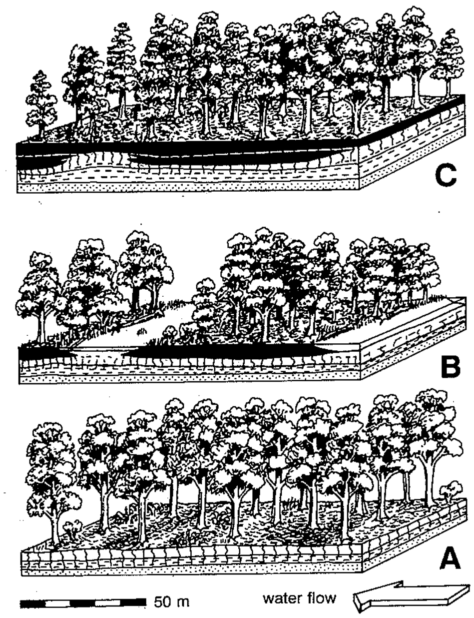


Figure 11. String bog (aapamire) model for the formation of stone-rolls in the floor of the Bulli and Wongawilli Coals.

Coal-bearing paleosols form by the accumulation of woody peat, which is commonly constrained to 0.5 to 1 mm/yr (Retallack, 1990, 1997b; Franzén, 1994). Seam splits in the Victoria Tunnel Seam of the Newcastle Coal Measures indicate compaction of the peat to 13% of its former thickness by overlying sediment (Diessel, 1992). This is a common value for peat to coal compaction of bituminous coals (Ryer and Langer, 1980). The duration of peat accumulation of Birdie paleosols can be estimated from their reconstructed thickness. An additional 1 k.y. should be added for Birdie paleosols because of their Inceptisol-like underclays, which have fossil roots and clay skins from a period of soil formation before the peat accumulated. These values, and a modest shale accumulation rate of 0.5 mm/yr based on varve thickness in the Murrays Run lacustrine sequence, can be used to compute sediment accumulation rates for different boundary sequences. In every boundary sequence examined (Figs. 3 and 4), sediment accumulation rate was faster in earliest Triassic time than in latest Permian time (Table 3).

A quickened pace of sedimentation also is apparent from a long-term sedimentation rate estimated from radiometric dating. The maximum thickness from the base of the Narrabeen Group to the base of the Minchinbury Sandstone of the Wianamatta Group is 1157 m (Uren, 1974; Stroud, 1974; Herbert, 1980). This thickness represents no more than 7.2 m.y. from the Permian-Triassic boundary at 250 ± 0.3 Ma (Renne et al., 1995) to the early Middle Triassic (Anisian) advent of *Dicroidium odontopteroides* in the Minchinbury Sandstone, a biostratigraphic event accomplished in New Zealand by at least 242.8 ± 0.6 Ma (Retallack, 1980; Retallack et al., 1993). These estimates give a long-term Early Triassic sedimentation rate of 0.16 mm/yr. In the latest Permian, the Thornton Claystone in the roof of the Big Ben Coal some 835 m below the top of the Newcastle and Tomago Coal Measures (Menzies, 1974) has been dated as 266.1 ± 0.4 Ma

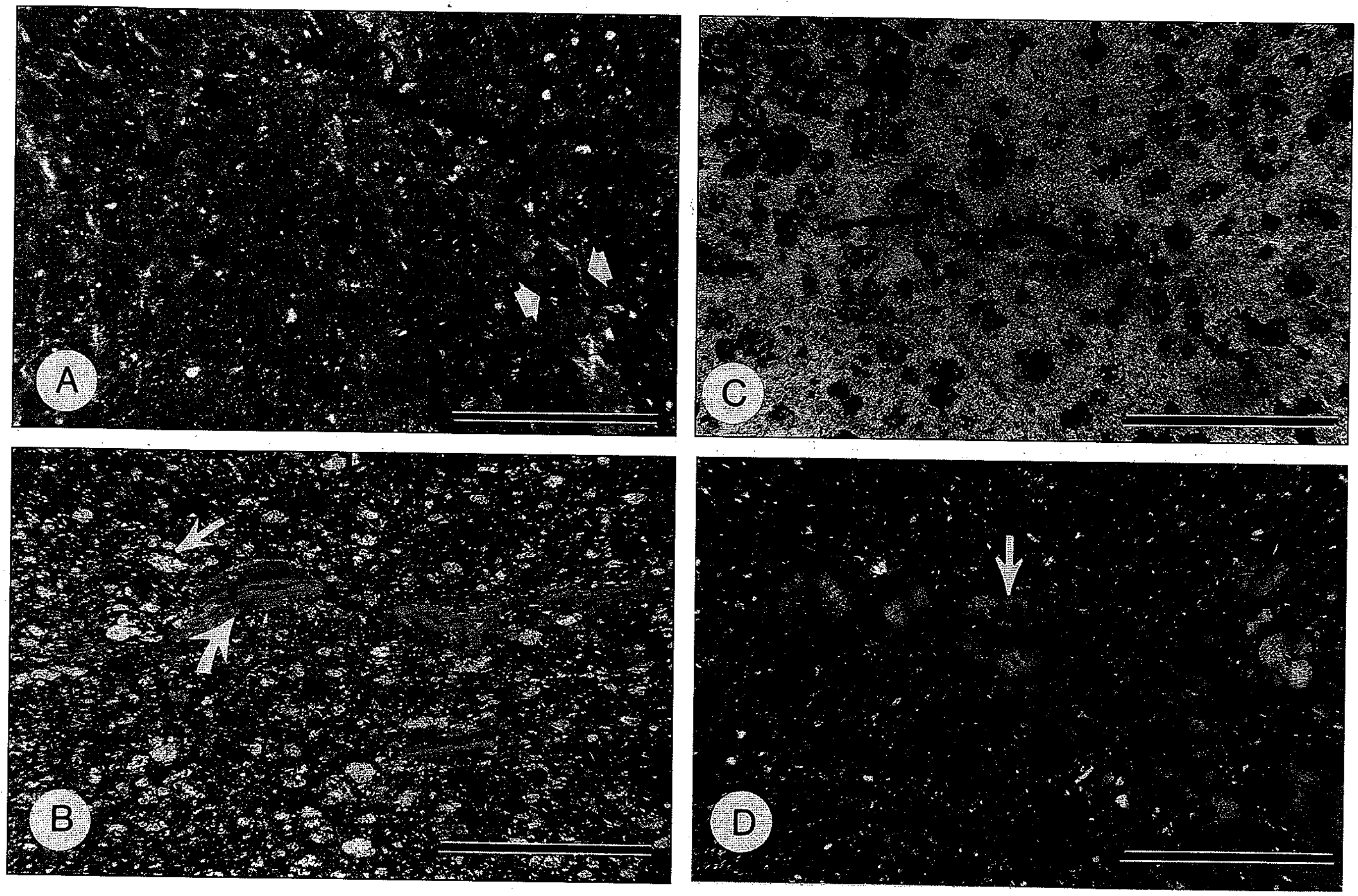


Figure 12. Photomicrographs of Early Triassic paleosols of the Sydney Basin, New South Wales. (A) Root trace (between wide white arrows), clay skins (light streaks), and clinobimasepic plasmic fabric under crossed nicols, from the subsurface (Bw) of type Wollemi paleosol at Coxs Gap (R1623). (B) Claystone breccia including clasts of pedogenic clay skins (above wide white arrow) and clasts of highly birefringent (omnisepic) soil claystone (at long white arrow) from C horizon of type Wybung paleosol at Wybung Head under plane light (R1667). (C) Ferruginized sphaerosiderite (microspherulitic siderite here stained opaque) under plane light, from parent material of type Wollemi paleosol at Coxs Gap (R1625). (D) Sphaerosiderite under crossed nicols (a clear example with cross extinction below white arrow), from the subsurface (Bg) horizon of the Wybung paleosol at Coalcliff (R1686). Scale bars =1 mm.

(Gulson et al., 1990), for a long-term Late Permian (Tatarian) sediment accumulation rate of 0.055 mm/yr. Comparable Late Permian rates can be calculated from preserved thickness of latest Permian rock above the 258 Ma Gerringong Volcanics and top of the Kiaman paleomagnetic superchron (Veevers et al., 1994a) or 256 Ma Awaba Tuff (Gulson et al., 1990). Also comparable are rates using a Permian-Triassic boundary age of 251 Ma (Claoué-Long et al., 1991). The order of magnitude difference between long-term radiometric and short-term paleosol-based rates is typical (Retallack, 1984), because of high short-term variability in sediment accumulation rate. Much of the missing time is in sequences of well-developed paleosols such as those of the late Early Triassic Bald Hill Claystone (Retallack, 1977a, 1977b, 1997c). Nevertheless, both radiometric and pedogenic estimates show acceleration of sedimentation rate at the Permian-Triassic boundary.

CESSATION OF COAL FORMATION

An obvious difference between latest Permian and earliest Triassic rocks in the Sydney Basin is the change from extensive peat swamps to non-peaty flood plains (Figs. 13 and 14). The lack of coal in earliest Triassic strata is especially anomalous because, as already noted, paleosols of that age indicate a humid, cool climate (from degree of weathering indicated by clay skins and high alumina/bases and barium/strontium ratios), waterlogged bottomlands (from carbonaceous and sideritic composition and high ferrous/ferric iron ratios), and also rates of sediment accumulation suitable for peat preservation (Table 3). There is similar evidence for humid paleoclimate, waterlogged bottomlands, and sediment accumulation rates in late Early Triassic paleosols of the upper

TABLE 3. SEDIMENT ACCUMULATION RATES (mm/yr^{-1}) ESTIMATED FROM PALEOSOL DEVELOPMENT

Sequence	Coalcliff	Bunnerong	Wybung Head	Murrays Run	Coxs Gap
Early Triassic	1.35	1.26	1.36	0.42	0.82
Late Permian	0.45	0.23	0.75	0.35	0.4

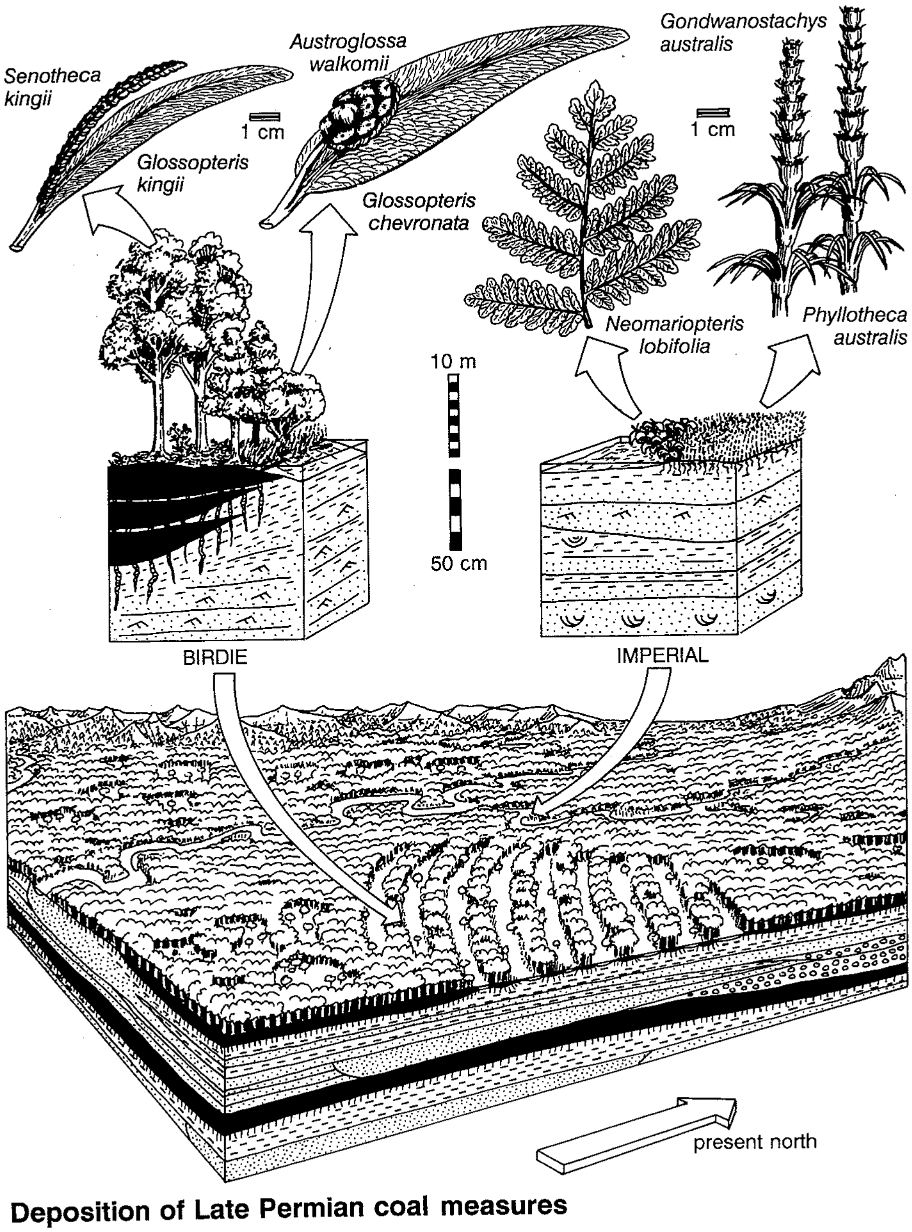


Figure 13. Reconstruction of latest Permian soils and vegetation of the Sydney Basin. Plants and vegetation are shown at different scales than soils and alluvial architecture, for which Figure 5 is lithological key.

Narrabeen Group of the Sydney Basin (Retallack, 1997c). Thus, the lack of coal cannot be due to low base level (Faure et al., 1995) or dry climate (Worsley et al., 1994). Nor is lack of coal merely a local phenomenon. No coal has been found anywhere in the world in Early Triassic rocks, representing the longest (about 6 m.y.) hiatus in coal formation since Devonian time (Faure et al., 1995; Retallack et al., 1996). The global coal gap also may be reflected in carbon isotopic ($\delta^{13}C$) data (Morante, 1996) and related to extinction of the peat-forming flora (Retallack et al., 1996).

MASS EXTINCTION OF PLANTS

Both fossil soils and plants indicate a profound change in the nature of vegetation at the Permian-Triassic boundary in the Sydney Basin (Retallack, 1995). As already outlined, latest Permian paleosols indicate swamp forests of cold-temperate to marginally frigid climate. Earliest Triassic paleosols, however, indicate forests adapted to low-nutrient clayey soils. The alkali and alkaline earth content of these paleosols is low despite partial restoration by local cementation and illitization during burial. A variety of Early Triassic forests are indicated by the variation within the paleosols of fossil root penetration (tabular versus deeply reaching). Also revealing are color (red versus gray) and oxidation state (FeO/Fe2O_3 molecular ratios), neither of which is affected by burial reddening or local burial gleization around root traces. From these indications some of these forested soils were well drained (Wollemi pedotype), while others were slowly drained (Cox) to poorly drained (Bongon, Wybung). The difference between Permian and Triassic paleosols may reflect in part the plant extinction across the boundary (Retallack, 1995).

These general impressions from the nature of root traces and paleosol profiles can be supplemented with evidence from fossil plants in some of the paleosols to add details of landscape ecology to the Permian-Triassic floral change in the Sydney Basin. Latest Permian *Glossopteris*-dominated assemblages are found within coal of Birdie paleosols, stratigraphically as high as 17 cm from the top of the Bulli Coal (Australian Museum specimen AMF91467–91468). Fossil plants collected in the underclay to the Bulli Coal at Coalcliff (AMF95387–95395) included not only *Glossopteris* leaves, but a variety of the distinctive fructifications of these plants: *Senotheca kingi* (McLoughlin, 1990a), *Squamella australis* (White, 1978), and *Austroglossa walkomii* (Holmes, 1974, 1995). Also found in coaly paleosols of the Birdie pedotype were roots of glossopterids, called *Vertebraria australis* McCoy (following Schopf, 1982, rather than McLoughlin, 1992). Only horsetail stems (*Paracalamites australis* Rigby 1969) were found in very weakly developed Imperial paleosols at Coalcliff. The horsetails can be interpreted as herbaceous vegetation early in ecological succession of disturbed ground. Other Imperial paleosols in the Murrays Run bore included better preserved herbaceous vegetation of horsetails (*Phyllotheca australis* Brongniart, Townrow, 1956) and ferns (*Neomariopteris lobifolia* Maithy, McLoughlin, 1992).

Almost all (97%) of the 33 Permian fossil leaf species described from the Sydney Basin were extinct by earliest Triassic time when no more than eight Triassic species based on leaves are known (Retallack, 1995). Claims for survival of *Glossopteris* into earliest Triassic rocks in the Sydney Basin cannot be substantiated. Some Middle Triassic leaves have been referred to *Glossopteris* (Holmes, 1992). Because their reproductive structures are unknown and their cuticles differ, the Triassic leaves are better placed in *Gontriglossa* (Anderson and Anderson, 1989). A large slab with beautifully preserved *Glossopteris* was mislabeled as being from the roof of the Bulli Seam (Diessel, 1992, fig. 6.7): it is actually from the roof of the Borehole Seam (C. F. K. Diessel, 1995, personal commun.). The stratigraphic level of the floral extinction is superbly exposed in the roof of underground coal workings of the Bulli Coal throughout the Southern Coalfield, in cliffs and collieries of the Blue Mountains, and in sea cliffs south of Newcastle: it is neither an artifact of collection nor an artifact of preservation (Retallack, 1980). The floral extinction also has been recognized in palynofloras in the Sydney Basin and throughout Australia (Helby et al., 1987) at the same level as a pronounced isotopic excursion in carbon composition of organic matter (Morante et al., 1994; Morante, 1996).

By earliest Triassic time a new flora had appeared. At Coxs Gap, fossil plants from a gray paleosol (Frazer pedotype) only 0.8 m above the last coal (AMF91440–91452) include abundant fronds of the seed fern *Dicroidium callipteroides* (White, 1984), with less common marattiaceous ferns (*Merianopteris* sp., *Cladophlebis carnei* Holmes and Ash 1979) and horsetails (*Paracalamites australis* Rigby 1969). Comparable plant assemblages are found within weakly developed paleosols in the basal Narrabeen Group at numerous other localities (Retallack, 1980): Bulli, South Bulli, Darkes Forest, Bellambi, Nattai, and Oakdale Collieries, Burragorang Valley, north Coalcliff Beach, and Frazer Inlet (Fig. 1). In Darkes Forest and South Bulli mines, maps of the sandstone paleochannels around plant localities in the roof of the seam show that this vegetation dominated by seed ferns formed an early successional scrub of scroll bars and levees of meandering streams (AMF91461–91466 and others discussed by Retallack, 1980).

Vegetation of better developed and more oxidized Early Triassic paleosols was dominated by conifers that were probably evergreen (Retallack, 1977b, 1980). This is indicated by a Kerrabee paleosol 13.6 m above the coal at Coxs Gap, which yielded mainly *Voltziopsis wolganensis* (Townrow, 1967) foliar spurs and cones (AMF91457–91459). Also found was the isoetalean lycopsid *Isoetes beestonii* Retallack (1997a; AMF91460) and a fern fragment of *Cladophlebis carnei* (Holmes and Ash, 1979; AMF91457). These also are common elements of the fossil flora of the basal Narrabeen Group elsewhere (Retallack, 1980, 1997a): including Rylstone, Newnes, and Bulli and Bellambi Collieries (Fig. 1). The quillwort (*Isoetes beestonii*) also forms monodominant even-aged stands in life position within lacustrine shales and can be regarded as an aquatic pioneering plant, like living quillworts (Retallack, 1997a). Living quillworts and conifers are both well-suited to nutrient-poor (oligotrophic) soils (Retallack, 1995, 1997a). The flora of the basal Narrabeen Group is very low in diversity compared with that of later Early Triassic time (Retallack, 1980, 1995, 1997a).

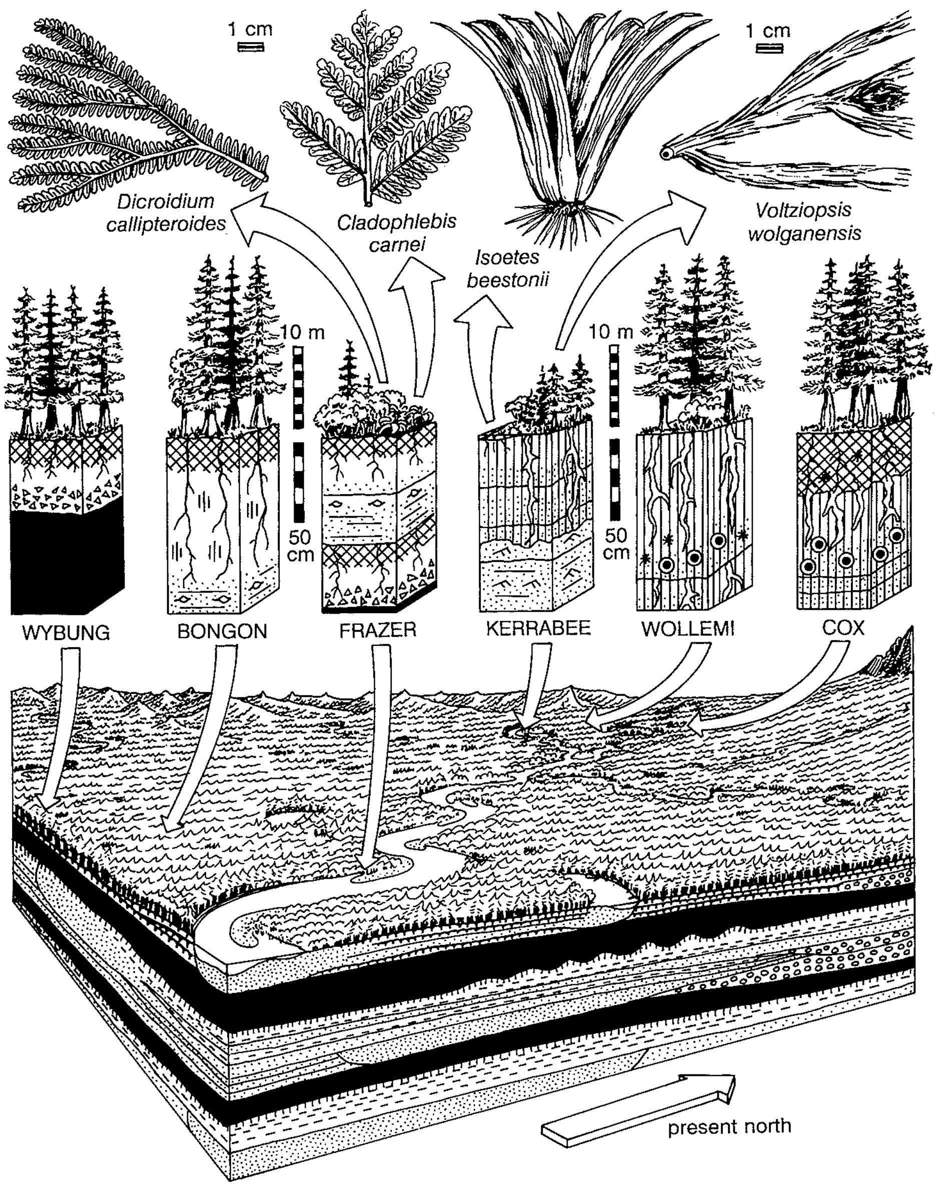


Figure 14. Reconstruction of earliest Triassic soils and vegetation of the Sydney Basin. Plants and vegetation are shown at different scales than soils and alluvial architecture, for which Figure 5 is lithological key.

MASS EXTINCTION OF ANIMALS

The Sydney Basin has a poor record of changes in animal life across the Permian-Triassic boundary. One reason for this may be the nature of the paleosols, which are noncalcareous and moderately oxidized, and so unsuitable for preservation of bones, shells, or organic carapaces (Retallack, 1984). Nevertheless, there are a variety of fossil tracks and burrows. Underclays of latest Permian coaly paleosols include a variety of burrows with backfill structures. Only a few simple vertical burrows with sandy fill, similar to *Skolithus* were seen in earliest Triassic paleosols (Bongon and Frazer pedotypes). A greater diversity of trace fossils has been found in paleosols of the late Early Triassic upper Narrabeen Group (Retallack, 1977a, 1995, 1997c: Naing, 1993).

Fossil insects are known from only a few localities in the Sydney Basin, but include as many as 144 species from the Late Permian, 1 from the earliest Triassic, 15 from the early Middle Triassic, and 12 from the late Middle Triassic strata. Similarly there are 25 known species of Late Permian conchostracans, but swarms of only 4 species in Early and Middle Triassic rocks (Retallack, 1995).

A well-preserved trackway from laminated earliest Triassic siltstones 25–50 cm above the coal in Bellambi Colliery is similar to that of *Lystrosaurus* (Retallack, 1996), and a variety of tracks are now known from the late Early Triassic upper Narrabeen Group (Naing, 1993). The fossil record of body fossils is biased toward aquatic assemblages of fish and amphibians, with little indication of latest Permian and Early Triassic land vertebrates in the Sydney Basin (Warren, 1991; Molnar, 1991).

ABRUPTNESS OF THE EXTINCTIONS

Estimates of time for formation of the paleosols also can be used to assess temporal resolution of the various measured sections. None of the sections have thick strongly developed paleosols that could be construed as evidence of a large time gap. Conglomerates of the Wybung Head section indicate very high sedimentation rates, which is strongly correlated with high temporal completeness (Retallack, 1984). Weakly developed paleosols at the Permian-Triassic boundary at Wybung Head and in the Murrays Run bore indicate breaks in sedimentation of no more than about 6 k.y., by comparison with well-dated Holocene soils (Ciolkosc et al., 1990; Tonkin and Basher, 1990). Additional evidence for temporal completeness comes from palynology. Boreholes near Wybung Head have preserved transient abundances of acritarchs, fungi, and ferns (Grebe, 1970) found in other parts of the world within about 50 k.y. of the boundary (Eshet et al., 1995; Visscher et al., 1996). The uppermost coal measures also contain glossopterid reproductive structures comparable to those found in the latest Permian rocks of the Bowen basin of Queensland (McLoughlin, 1990a, 1990b, 1992, 1994), and the Estcourt Formation of South Africa (Anderson and Anderson, 1985). The Sydney Basin also has the youngest known Permian palynozone of *Microreticulatisporites bitriangularis* (Price, 1983). *Playfordiaspora crenulata* has been thought to represent an additional latest Permian palynozone within the lowest 7 m of the *Protohaplyxoypinus microcorpus* zone in the Rewan Formation of the Bowen basin (Foster, 1982). However, the *crenulata* palynozone is earliest Triassic in age, because it has been correlated with the uppermost Chhidru Formation of Pakistan, which contains Early Triassic conodonts (Sweet, 1992) and lies above the $\delta^{13}C_{carb}$ crash that marks the Permian-Triassic boundary in Pakistan and elsewhere (Baud et al., 1989). Palynomorphs of the *crenulata* zone are also widespread in the basal *microcorpus* zone in the Sydney Basin (Price, 1983), but the zone has not been widely recognized in Australia (Helby et al., 1987). Thus, the Sydney Basin has the oldest Triassic and youngest Permian fossil plants and pollen, as well as transient fungal and fern spikes at the Permian-Triassic boundary, indicating a record that is as complete as any others known.

Further evidence comes from the continuous rather than abrupt decline of carbon isotopic composition ($\delta^{13}C_{org}$) of kerogen at the boundary in the Murrays Run bore in the northern Sydney Basin (Morante, 1996). Comparably gradational isotopic declines are also found in the most complete sections analyzed in this way elsewhere (Holser and Magaritz, 1987; Holser et al., 1991; Wang et al., 1994). Such complete isotopic records can be contrasted with the abrupt truncation between Permian and Triassic isotopic composition found in the disconformable sections of the Canning basin of Western Australia (Morante et al., 1994; Morante, 1996).

Finally, there is evidence from borehole correlations throughout the Sydney Basin (Arditto, 1991; Herbert, 1995, 1997a, 1997b) that the youngest Permian and oldest Triassic rocks in the basin are in the region of Wybung Head and Murrays Run bore. The uppermost Permian coal at Wybung Head can be correlated with carbonaceous shale containing the Permian plant fossils *Glossopteris* and *Neomariopteris* in the Murrays Run bore. There is onlap of about six fluvial cycles (parasequences) of paleochannel sandstone grading up to clayey paleosols, representing a hiatus of about 600 k.y. in the southern Sydney Basin. There is an even more marked disconform-ity toward the northwest near Coxs Gap, where the time gap may be as much as 1 m.y. A comparable hiatus is likely around the Lochinvar dome, which was tectonically active during and before this time (Herbert, 1997a). These interpretations are compatible with absence of the fungal spike in the southern coalfield and of the *Dicroidium callipteroides* zone around the Lochinvar dome. The duration of the Permian-Triassic discon-formity thus varies with location within the basin, from as much as 0.6 to 1 m.y. in the southern and western onlapping margins of the basin, to as little as 6–50 k.y. near Murrays Run and Wybung Head in the northern depocenter to the basin.

CONSTANT DEPOSITIONAL SETTING

Profound changes in vegetation, climate, sedimentation rate and other remarkable event markers could perhaps be explained by dramatic uplift or sea-level change. Surprisingly, major features of the depositional basin did not change across the Permian-Triassic boundary (Cowan, 1993). The Sydney Basin remained a nonmarine foreland depression, with a thick sequence accumulating to the southwest of the New England Fold Belt to the northeast, and thinner sequences accumulating on the stable ramp of the Lachlan Fold Belt to the south and west. Growth of structures such as the Lochinvar Dome to the north continued across the boundary (Herbert, 1997a), but no dramatic acceleration of that uplift is indicated. To the east, in what is now the offshore Sydney Basin, was an extinct volcanic arc of latite (shoshonite) volcanoes (Bradley, 1993).

Drainage of this vast alluvial plain was generally to the southeast from the northern upland, with less prominent eastward and westward flow from basin margins, both during Late Permian and earliest Triassic time (Diessel et al., 1967; Ward, 1972: Bowman, 1980; Cowan, 1993). There were marine incursions into the basin earlier in Permian time, but the latest Permian coals have very low boron and sulfur content, and low amounts of durain, indicating no remaining influence of seawater (Diessel, 1992). There is, however, evidence for earliest Triassic sea level rise and subsequent Early Triassic inundations from the local abundance of acritarchs in palynological preparations (Grebe, 1970; Helby et al., 1987).

Thick, laterally extensive conglomeratic paleochannels in latest Permian and earliest Triassic strata at Coxs Gap and Wybung Head are similar to those of alluvial fans (Diessel, 1992). They contain numerous rounded clasts of chert, schist, and vein quartz from the New England Fold Belt, which would then have been a substantial mountain range not far to the north and east. Bed thickness and clasts in these conglomerates are smaller in the Triassic strata than in the Permian strata within each section, indicating a lack of tectonic uplift coincident with the boundary (Figs. 3 and 4). The basal Triassic conglomerates also lack red, chemically oxidized paleosols, so these alluvial fans were not building above base level. In the Murrays Run and Bunnerong bores and on the south coast near Coalcliff, clayey paleosols are more common and paleochannels are mainly sandstone, with fining-upward cycles typical of meandering streams (Diessel, 1992). Maps of anastomosing earliest Triassic sandstone paleochannels in the roof of the Bulli Coal have been interpreted as evidence for braided streams, but their high sinuosity and associated features such as scroll bars are evidence that the anastomosing pattern is an artifact of avulsion of meandering streams (Retallack, 1980). Geomorphic setting varied over this large basin, but within each region examined there are similarities in alluvial architecture across the Permian-Triassic boundary.

Nor is there petrographic indication of tectonic events across the Permian-Triassic boundary, with the exception of the thin (5–20 cm) geochemically anomalous breccia beds at the boundary already discussed. All the sandstones have predominantly lithic grains, with minor amounts of quartz, feldspar, siderite, and clay (Loughnan, 1963, 1966). Clays of the sequence are mainly kaolinite and illite, with smectite and analcite in tuffaceous beds more common to the north than in the southern and western Sydney Basin (Loughnan, 1963, 1966). There is no flood of metamorphic and volcanic clasts that would be expected from tectonic uplift of the source terranes or changes in local base level.

EARLIEST TRIASSIC GREENHOUSE

The Sydney Basin shows a remarkable concatenation of events across the Permian-Triassic boundary. High latitude latest Permian and earliest Triassic paleosols of the Sydney Basin demonstrate a warmer paleoclimate following end-Permian mass extinction of plants and animals. As already discussed, the transition from coal measures to red-green mottled paleosols was not due to long (millions of years) erosional disconformity, markedly drier or wetter climate, tectonic uplift, or falling sea level. Furthermore, the cessation of peat accumulation was global. No Early Triassic coal is known anywhere in the world (Retallack et al., 1996). Much of this carbon probably vanished into thin air by burning, decay, or release of methane clathrates at the inception of the Mesozoic greenhouse (Erwin, 1993, 1994), as indicated by the synchronous carbon isotopic shift in nonmarine kerogen (Morante et al., 1994; Morante and Herbert, 1994; Morante, 1996). A similar atmospheric shift to a CO_2 or CH_4 greenhouse is indicated by oceanic decrease in $\delta^{13}C$ values of carbonate and organic matter (Holser and Magaritz, 1987), although the coeval sulfur isotopic excursion may have been larger than anticipated in their modeling (Worden et al., 1997). Models of atmospheric change based on estimated carbon budgets show a decline of oxygen from an Early Permian high of 35% to an Early Triassic low of 13% (present level is 21%), and a concomitant rise of carbon dioxide from 0.03% (near present level) to 0.15% (5 times present atmospheric level) in Early Triassic time (Graham et al., 1995). Evidence from $\delta^{18}O$ values of carbonates in Austria indicates a 6–11 °C rise in temperature across the Permian-Triassic boundary (Magaritz and Holser, 1991; Holser et al., 1991). Evidence from paleosols presented here indicates that the onset of the early Mesozoic greenhouse was a marked climatic shift in formerly high latitude areas such as the Sydney Basin.

From this perspective a plausible explanation for accelerated rates of sediment accumulation and the distinctive claystone breccias of the earliest Triassic is a profound disruption of plant communities. Transient abundances of fungi and fern spores at the Permian-Triassic boundary can be interpreted as evidence of ecosystem destruction and rotting followed by recolonization by herbaceous plants (Eshet et al., 1995; Visscher et al., 1996). A deforested landscape would have been prone to catastrophic soil erosion. The low-diversity, oligotrophic flora of earliest Triassic time (Retallack, 1995, 1997a) also had low primary productivity, judging from the low $\delta^{13}C$ of organic matter (Morante et al., 1994; Morante, 1996). This surviving flora may have been less effective in landscape stabilization than preexisting *Glossopteris* forests. Increased rates of sediment accumulation in Early Triassic time may thus be due to a less effective new regime of soil binding by surviving vegetation.

Widespread deforestation followed by soil erosion would produce the claystone breccias, accelerated rates of sedimentation, and oxidation of land plants by burning or decay seen in the Sydney Basin. This could lead to eutrophication, swarms of microplankton, anoxia and extinctions in nearby lakes and oceans. Swarms of acritarchs (Balme, 1969; Helby et al., 1987) and a proliferation of fungi (Eshet et al., 1995; Visscher et al., 1996) are now recognized worldwide in earliest Triassic rocks. Oceanic anoxia (Wignall and Hallam, 1993; Kakuwa, 1996; Wignall and Twitchett, 1996; Isozaki, 1997), excess bicarbonate (Kempe and Kazmierczak, 1994; Knoll et al., 1996), low pH (Liu and Schmitt, 1992), and acid rain (Hsü and McKenzie, 1990; Conaghan et al., 1994) have all been blamed for the Permian-Triassic extinctions in the ocean, but may instead be consequences of widespread deforestation. In any case, it is difficult to understand how these oceanic mechanisms could have such profound effects on land, either in the Sydney Basin discussed here, or in the Karoo Basin of South Africa (Smith, 1995; MacLeod et al., 1997).

If these are consequences, then the ultimate cause of the extinctions, deforestation, and geochemical anomalies may have been bolide impact (Hsü and McKenzie, 1990; Retallack et al., 1996; Kerr, 1996; Gorter, 1996), methane release from permafrost and marine continental shelves (Erwin, 1993, 1994; Morante, 1996), or voluminous volcanic eruptions of the Siberian Traps (Conaghan et al., 1994; Renne et al., 1995; Gurevitch et al., 1995). There are problems with these explanations also. The light carbon of the isotopic excursions cannot have come from volcanic eruptions and persisted too long into Triassic time to be caused by catastrophic release of CH_4 or CO_2 at the boundary (Holser, 1997). Possible impact craters include a 500-km-diameter structure in the offshore Canning basin of Western Australia dated by conventional K/Ar at 253 ± 5 Ma, known only from a preliminary report (Gorter, 1996). The 40-km-diameter Araguainha crater of Brazil is now known to be significantly younger than the Permian-Triassic boundary (Hammerschmidt and Engelhardt, 1995, Engelhardt et al., 1994). Shocked quartz at the Permian-Triassic boundary in Antarctica and Australia is much smaller and rarer than at the Cretaceous-Tertiary boundary, and within a redeposited soil deposit rather than an impact breccia (Kerr, 1996; Retallack et al., 1996a). Most microspherules from the Chinese boundary beds appear volcanic, and only a few bearing chrome and spinel are plausibly meteoritic (Yang et al., 1995). Iridium anomalies at the Permian-Triassic boundary in Armenia (Alekseev et al., 1983), Italy (Oddone and Vanucci, 1988), Austria (Holser et al., 1991), and China (Zhou and Kyte, 1988; Xu and Yan, 1993) are faint to insignificant, or otherwise problematic (Holser et al., 1991). The discovery of a positive europium anomaly at one boundary section in India can be interpreted as evidence of an iridium-poor eucritic impactor (Bhandari et al., 1992). This europium anomaly has not yet been found in China (Zhou and Kyte, 1988), where other trace element studies indicate only about 30% meteoritic contribution to the largely volcanic boundary beds (Chai et al., 1992). There are no easy answers yet.

It is unlikely that debate about ultimate causes of the Permian-Triassic extinctions will be resolved any faster than debate about extraterrestrial causes for the Cretaceous-Tertiary extinctions. However, the Sydney Basin includes important local sections for understanding the Permian-Triassic boundary. Paleosols of latest Permian and earliest Triassic age provide several pieces of the puzzle formerly unattainable. The most striking of these new discoveries is the warming at the Permian-Triassic boundary, here termed a postapocalyptic greenhouse.

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