WEATHERING HISTORY, LANDSCAPE EVOLUTION AND IMPLICATIONS FOR EXPLORATION

R. R. Anand

CRC LEME c/o CSIRO Exploration and Mining, 26 Dick Perry Avenue, Kensington, WA 6152

INTRODUCTION

The landscape of Australia is unique. Parts of the continent have been emergent and exposed to sub-aerial conditions for over one billion years (Figure 1); elsewhere, there has been marine sedimentation to the Holocene. Many of the landforms have their origins 300 Ma ago, when the break-up of Gondwana began. During this long period, the rocks that form the land surface have experienced a wide range of climates, including glacial, temperate and tropical, and humid to arid, with concurrent tectonic events, such as uplift, warping and continental break-up. These caused great variations in the chemical and physical environment that have resulted in intensive weathering and alteration of the exposed rocks. Because of the tectonic stability of much of the continent during the Phanerozoic, the weathered mantle, the *regolith,* has been preserved rather than eroded, and is present as a widespread cover, 20 to >100 m thick over much of the continent. Generally, the regolith thus records complex interactions between weathering and erosive processes. Regolith not only includes weathered *in situ* materials but also transported materials of variable thickness, including complex palaeochannel systems.[1]

From a geochemical exploration perspective, weathering causes the destruction of primary ore deposits and the dispersion of ore and pathfinder elements in the surrounding regolith. Conversely, it may also result in the supergene enrichment of some deposits and promote the formation of secondary orebodies. To understand the history and potential mechanisms and pathways of migration of ore and pathfinder elements in regolith, it is necessary to unravel the complex superposition of events that may have occurred during regolith-landscape evolution. Similarly, an understanding of the regolith is applicable to a multitude of environmental and economic problems in our society, perhaps best exemplified, in Australia, by dryland salinity.

Regolith-landscape studies have a long history, dating back to the early scientific explorers in the mid-nineteenth century, although systematic scientific work dates mainly to early last century. Woodall (1993) and Smith (1996) have noted that, since 1980, much of the research related to the regolith has been driven by mineral exploration and has concentrated on developing exploration procedures appropriate for regolith-dominated terrains. In the last few years, there has been an increased emphasis on weathering geochronology, palaeoclimatology, 3D regolith architecture and understanding the processes of regolith formation.

This chapter provides a broad overview of the weathering history and evolution of Australian landscapes. The main objectives are to: (i) summarize the geological history of the Australian landscape; (ii) summarize the weathering and regolith-landform processes, and (iii) discuss the implications for exploration. A detailed synthesis of the geochronology of the Australian regolith is given by Pillans (2005).

PHYSICAL FEATURES

Except along the eastern seaboard, much of Australia has been tectonically stable since the Proterozoic. Erosion has progressively levelled large areas of the continent to low-lying plains, leaving a relatively flat landscape (Figures 2A and B). The Western Plateau consists of the main Archaean and Proterozoic structural domains where the landscape is largely plateaux and ranges. The Interior Lowlands consist of vast floodplains and low-relief landforms. The Eastern Uplands are more complex and younger, with elevated remnants of the Tasman Fold Belt. These uplands slope gently W to the Interior Lowlands and more steeply E to a very narrow coastal plain.

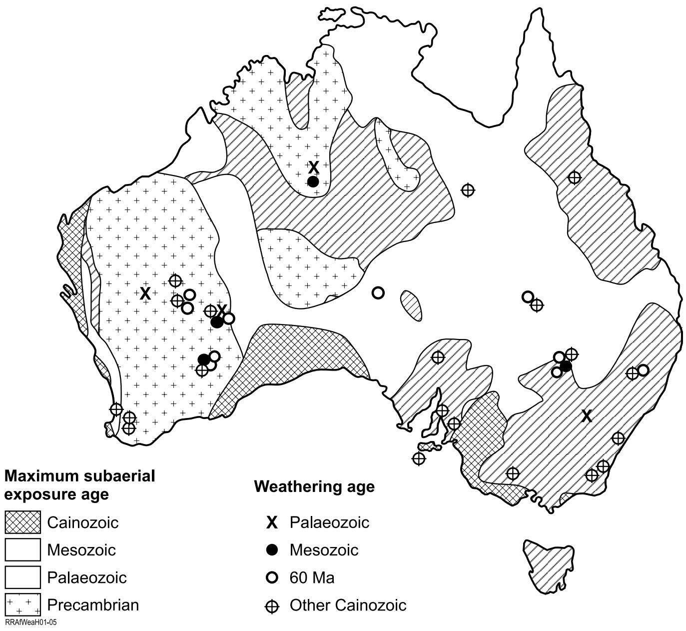

Figure 1. Minimum duration of subaerial exposure (data from BMR Palaeogeographic Group, 1990) and palaeomagnetic ages of the regolith (after Pillans, 2005).

The physiography has a profound influence on the orographic component of rainfall. The SW quadrant of the continent has a low rainfall of 200-400 mm pa (Climate, *this volume*), except in the low hills of the Darling Range in SW Western Australia. In contrast, the Eastern Uplands attract significant orographic rain (600-1000 mm pa). The central part of Australia, lacking any significant topography, is predominantly dry, whereas the N has a monsoonal influence. The total annual rainfall is generally highly variable, particularly over the arid interior, and in parts of the monsoonal north, where the irregular incidence of tropical cyclones may cause great variation in yearly totals. The temperature regime is essentially continental, with the high temperatures of the inland reaching almost to the coastline in some areas. Average annual air temperatures range from 29°C along the NW coast area, to 5°C in the alpine areas of the SE. In January, average maximum temperatures exceed 36°C over vast areas of the interior and the NW. In July, average minima fall below 6°C in areas S of the tropic and away from the coasts. Over 75% of the continent, average annual evaporation is high, exceeding 2000 mm, and rainfall does not exceed evaporation loss from a free water surface in any month of the year.

The native vegetation of Australia is dictated, both in form and distribution, more by rainfall than by any other physical factor, so that there is a general sequence of plant formations related to broad climatic regions (Vegetation Communities, *this volume*). Australian trees and shrubs are predominantly evergreen and their leaves are, in general, hard and tough. A few distinctively Australian genera, notably *Eucalyptus*, *Acacia*, and *Triodia* dominate and give a characteristic appearance to much of the vegetation. In contrast with many other similarly arid regions of the world, the lower rainfall areas of Australia support a high proportion of woody species and have no large indigenous succulents.

The physiography of the inland Australia, combined with the associated climatic features and immense areas of semi-arid and arid lands, is allied with a very poor external drainage system. Coordinated external drainage occurs only in the outer, peripheral zone (Figure 3) and, even here, flow is permanent only where there is high rainfall. Coordinated drainages also occur in the Murray-Darling Basin, which has an external

1Footnote: Regolith is the entire unconsolidated and secondarily recemented cover that overlies more coherent bedrock that has been formed by weathering, erosion, transport and/or deposition of older material. It includes fractured and weathered basement rocks, saprolites, soils, organic accumulations, glacial deposits, colluvium, alluvium, evaporitic sediments and aeolian deposits.

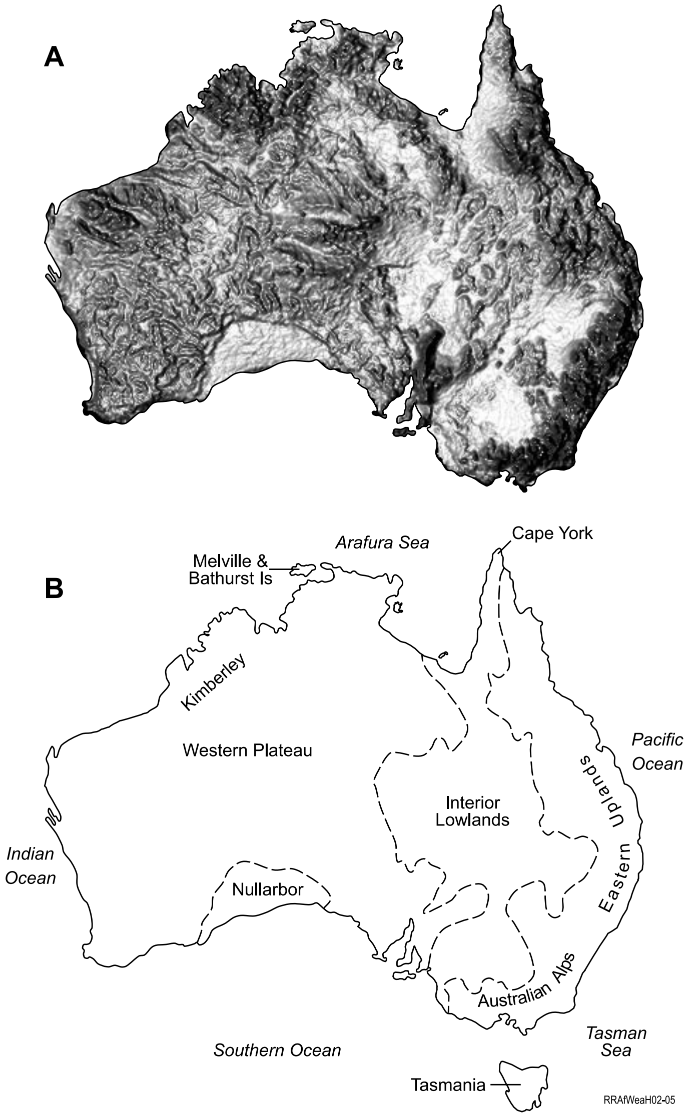

Figure 2. A. Digital elevation model with illumination from the NW (AUSLIG) showing the major physiogeographic regions. B. Subdivision of the continent into major physiographic regions (after Parkinson, 1988).

outlet for part of its total discharge. The Lake Eyre Basin, which has no external outlet, loses its low input by evaporation. The remainder of the country has either uncoordinated surface drainage or no surface drainage. In the former, there is brief connected flow only after rare, heavy rainfall. However, here, the drainage has sub-surface connections that follow ancient choked river channels, now shown by chains of playas and, as such, have outlets to the ocean. Short connected drainages occur in local areas of moderate relief, but these die out on the plains.

Lack of coordinated drainage and low topographic gradients severely limit the application of regional stream sediment geochemistry in much of central Australia. Where the drainage is internal and the evaporation rates high, solutes are concentrated in the groundwater which, in places is over ten times more saline than seawater.

GEOLOGICAL HISTORY AND PALAEOCLIMATES

A detailed account of Australia's geological history in the Phanerozoic is beyond the scope of this chapter. Instead, essential information is derived from existing reviews (Ollier, 1977; Kemp, 1978; White, 1994; Frakes, 1999; MacPhail, *in press)*. A great ice sheet covered about half of the Australian landmass in the Permian (Figure 4A). This provided a fresh start for landscape evolution, comparable to that of the Quaternary glaciation of the northern hemisphere. However, palaeomagnetic results indicate that significant weathering occurred during the Palaeozoic and despite substantial erosion, remnants of the regolith, formed during the Palaeozoic, have survived in places (Figure 1).

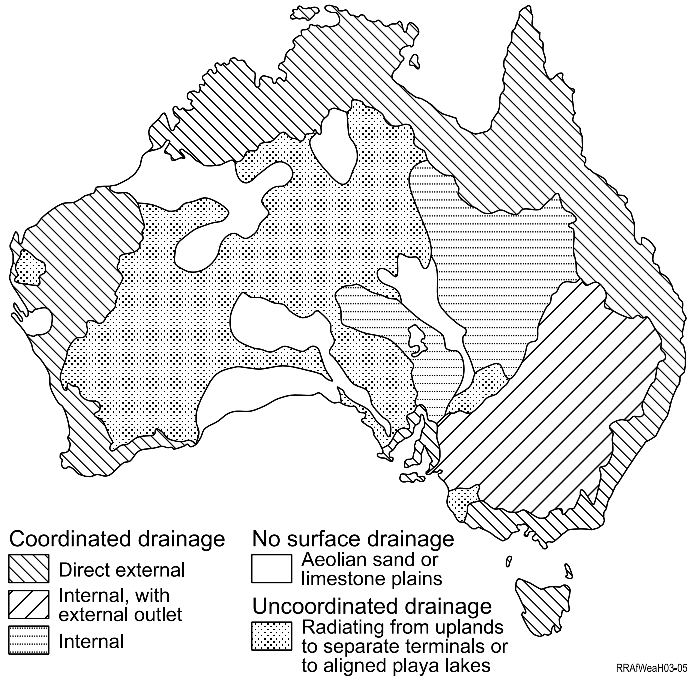

Figure 3. Drainage divisions of Australia (after Mabbutt, 1973).

A line of marine basins (the Tasman Geosyncline) ran along the eastern side of the continent from Tasmania to Queensland (Ollier, 1977). The cool to cold climates of the Early Permian, which supported glaciers on the eastern highlands gave way to warmer climates and the development of coal swamps. Accordingly, these basins were filled first by glacial sediments and then by marine or terrestrial sediments, including important coal deposits. Volcanic activity along the line of the basins also contributed sediment (Ollier, 1977).

Shallow water marine sediments were deposited in the Canning, Carnarvon and Perth basins after glacial deposits had accumulated within them. The present landsurface of the Yilgarn Craton is only a few metres below the essentially flat-lying, sub-Proterozoic unconformity, and the overall flatness of the landscape suggests that the plateau may represent a Proterozoic erosion surface (Daniels, 1975). The contribution from reworked Permian is unknown although undoubtedly important (Clarke, 1994). Alley and Clarke (1992) reported the presence of reworked Late Permian palynomorphs in Cretaceous sediments from the Great Australian Bight, indicating sediments of this age were present in the hinterland at the time of deposition, although none is now preserved.

During the Triassic, the continent was almost entirely emergent, with several basins where lake and river sediments accumulated (Figure 4B). The coal swamps had largely disappeared, with only small areas in South Australia, eastern Tasmania, NE New South Wales and adjacent SE Queensland. Volcanoes were active along the eastern continental margin, on the continental shelf and on land E of the major basins. Although Australia still occupied high latitudes, the climate appears to have been generally warm (Figure 5).

The area of sedimentation increased considerably during the Jurassic (Figure 4C). In eastern Australia, Jurassic deposits are almost exclusively non-marine, although marine sediments were deposited within the Canning and Perth Basins in Western Australia. Volcanism was very restricted. Doming, which preceded the separation of New Zealand from Australia in the first stages of the break-up of Gondwanaland, gave the first major uplift of the eastern highlands of Victoria (Wellman, 1974). At the same time, great thicknesses of dolerite were being emplaced as sills in Tasmania, in association with similar sets of movements - perhaps the precursor to separation of Australia and Antarctica. Australia remained in the high latitude humid belt during the Jurassic (Figure 5).

The plate tectonic movements, which were starting to break the super-continents, led to a global sea level rise in the Early Cretaceous. In the first 50 million years of the Cretaceous, the Tasman Depression

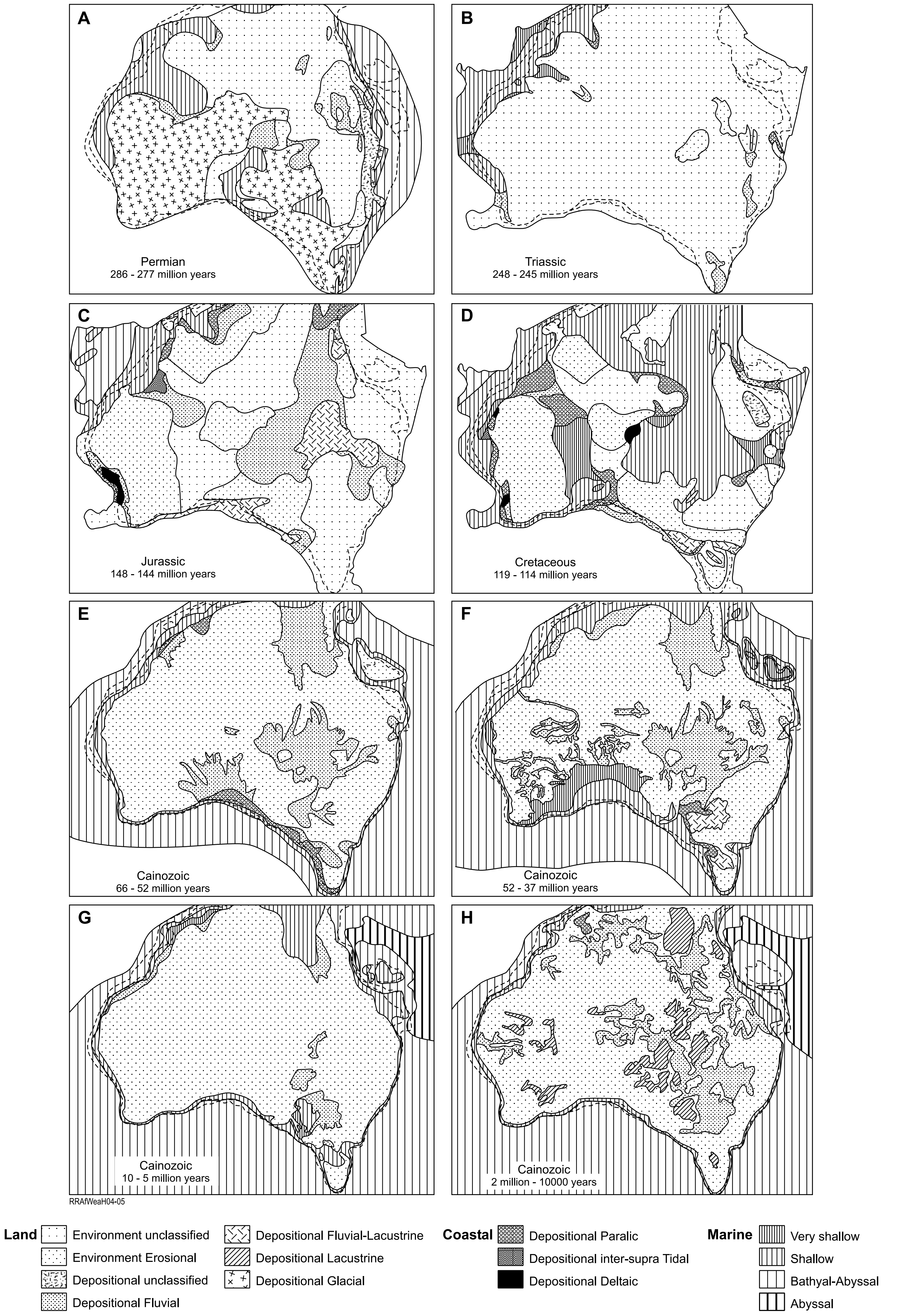

Figure 4. Paleogeographic maps of the Australian continent (Copyright Commonwealth of Australia) (Geoscience Australia, 2005).

and adjacent low-lying central and southern basins, as well as basins in Western Australia were flooded (White, 1994) and the vast waters of the Eromanga Sea divided the continent into three landmasses (Figure 4D). As a result, Cretaceous sediments outcrop or are buried by younger sediments over approximately one third of the present landmass. There are no known occurrences of Cretaceous sediments on the Yilgarn Craton. Lower Cretaceous sediments of the Madora Formation (Lowry, 1970) are widespread in the Officer and Eucla Basins, but do not appear to have been deposited beyond the current basin margins. By the Late Cretaceous, the sea had withdrawn from most of the continent and extensive river system developed in the region of the Great Artesian Basin. Elsewhere, the drainage systems that would persist through the Tertiary to the present day were also initiated. The climate in Australia was very warm and humid (Figure 5) during most of the Cretaceous, although there were cold spells at the beginning and end (White, 1994). Tropical and sub-tropical conditions extended much further N and S than at present, and there was a warm-temperate climate at the Poles.

The cumulative effects of planation during the Permian glaciation, widespread Early Cretaceous sedimentation, and emergence and erosion of much of the continental land mass from the mid Cretaceous (Figures 4E and F), meant that by the start of the Tertiary, Australia was already a remarkably flat continent (Ollier, 1977). This situation continued, but broad epeirogenic movements gave rise to some regional variety. A series of uplifts, collectively known as 'Kosciousko Uplifts', have been active intermittently since the Middle Miocene to Early Pliocene, primarily in the SE of the continent (Abele, 1976). Uplifts of the eastern highlands probably began on a small scale in the Eocene, peaked in the Miocene and have continued gradually and episodically ever since (Wellman, 1974). Ollier (1977) has noted that late tectonic movements in Victoria have been important in producing physiographic effects away from the highlands, in the form of major coastal inlets, in the outline of the Western Plains, and in supposed fault block topography in the SE. Uplift of the Flinders Ranges in South Australia probably commenced in the Late Miocene (Callen, 1977) and continued after this time. Subsidence of the Murray-Artesian depression and up-warping of the western half of the continent proceeded gradually and there was some faulting and block movement in marginal basins. Estimates of the thickness of material eroded from all or part of the Yilgarn Craton range from 400 m, or less, to over 5 km (Finkl and Fairbridge, 1979; van der Graaf, 1981; M.J. Killick, written communication, 1998). This topic is discussed further by Ananad and Paine (2002).

Marine Tertiary deposits are known from the Murray and Eucla basins in the S, the Carnarvon Basin in Western Australia and all around the continental shelf, notably on the Great Barrier Reef platform off the Queensland coast. The first cycle of Cainozoic marine transgressions along the southern margin of Australia (cycle 1) is not represented in the stratigraphic record of the onshore Eucla basin (Lowry, 1970; Hocking, 1990). In the western Eucla Basin, the shoreline remained on or below the present continental shelf throughout the Early Tertiary, even during the Late Palaeocene-Early Eocene when the sea levels were relatively high (de Broekert, 2002). The reason why the Late Palaeocene-Early Eocene high stands failed to penetrate the onshore part of the Eucla Basin is unclear, but is probably because subsidence of the Eucla Basin was slower in the W than the E. On the Yilgarn Craton to the W and Officer Basin to the N, the formative streams flowed within the base of an ancient system of broad, shallow valleys, termed 'primary valleys' by de Broekert, 2002. These were in existence by the earliest Cretaceous, transferring large quantities of terrigenous sediment to the evolving rift valley (Bight Basin and Eyre Sub-Basin) between Australia and Antarctica, which began to develop in the latest Middle Jurassic. It is important to note that these basin-fills are compositionally immature, containing abundant feldspars with lesser pyroxenes (Hill, 1991; Clarke and Alley, 1993), whereas the inset-valley fills comprise kaolinite and quartz, both of which are chemically stable.[2] Thus, it would appear that primary valleys were set within fresh basement, which had become deeply weathered by the time the inset-valleys were formed. Deep weathering within the primary valleys masked the influence of bedrock lithology and structure on the inset-valley incision and did much to prepare the materials from which the inset-valley fills were sourced. Incision of inset-valleys is most likely to have been caused by stream rejuvenation following epeirogenic uplift. Comparison with modern valleys formed by the climate change and a review of Tertiary climate, all indicate that a change in climate is unlikely to have caused the inset-valleys to form. Palynological evidence for the widespread occurrence of rainforest (Clarke, 1993, 1994a, 1994b) and the lack of major climatic events in the Early Tertiary argue particularly strongly against climate change as even having had a contributory role in inset-valley incision (de Broekert, 2002).

The age of inset-valley incision could be estimated by dating palaeosols, or similar near-surface alterations, formed beneath the gentle side-slopes of the inset-valleys during the course of their incision. Ferruginous lateritic residuum developed from basement rock occurs directly beneath some inset-valleys in the northern Yilgarn (Anand and Paine, 2002). However, these materials have not been dated, and in any case, there is some uncertainty as to whether they formed before, during, or after inset-valley incision. de Broekert (2002) concluded from his study in the Kalgoorlie region that, in the absence of a quantitative measure, the onset of inset-valley incision probably occurred in the Early Palaeocene to Middle Eocene.

The driving force for climate change during the Tertiary was Australia's separation from Antarctica, which began in the latest Early Cretaceous, coupled with global cooling which started in the latest Early Eocene and continues to the present (MacPhail, *in press*). Australia was 25° S of its present position at the beginning of the Tertiary and has since continued to move N through a series of climatic zones, which were themselves

2Footnote: The buried valleys are commonly referred to as 'palaeochannels'; herein, they are termed 'inset-valleys', to emphasise their subordinate and entrenched position within the bedrock surface of another system of much broader and subtly-defined 'primary-valleys' (after de Broekert, 2002).

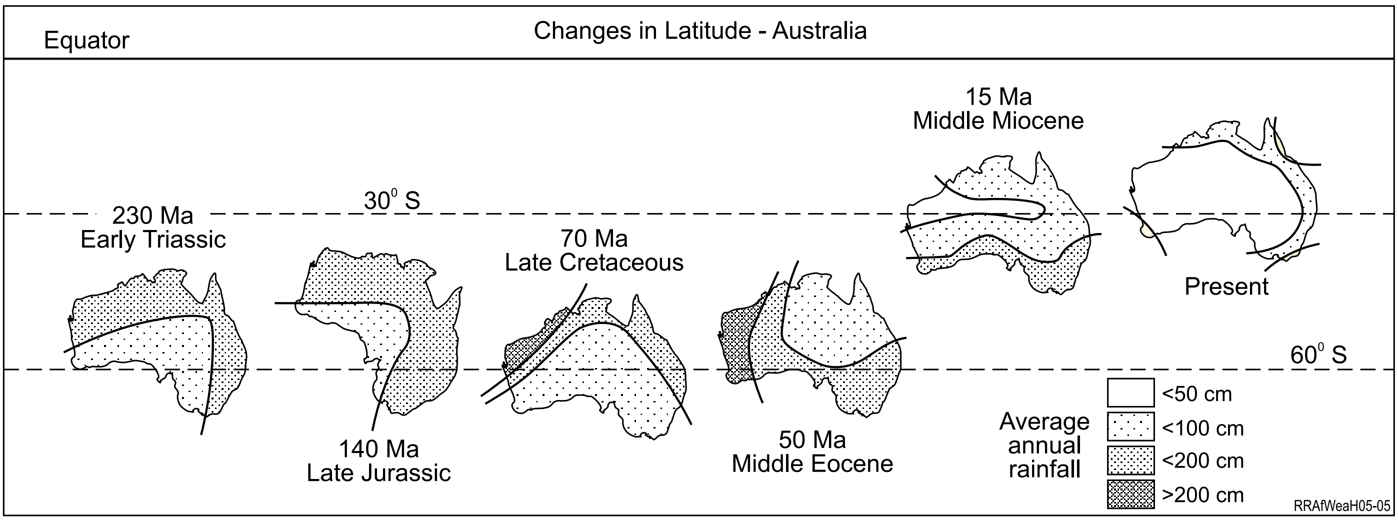

Figure 5. Palaeoclimate history of Australia (after Tardy and Roquin, 1998).

changing position over the Tertiary (Figure 5). In the Palaeocene and Eocene, the climate was warm and wet, and rainforest covered much of the land. During the Oligocene, the overall temperature became cooler and nearer to those at present (Frakes and Kemp, 1972), but became warm to hot again in the Early Miocene. By the Mid Miocene, the permanent southern ice cap formed and, thereafter, the cooling trend again became evident and subsequently drying of the Australian continent accelerated as the world cooled and the vegetation changed. Rainforests contracted, the centre of Australia became desert and the fauna progressively adapted to the changing environment. Only the SW and N margins currently have humid climates, respectively Mediterranean and savanna.

During the Quaternary (approximately the last 2 million years) there have been at least 20 oscillations from glacial to interglacial climates, caused, at least in part, by changes in thermal equilibrium resulting from variations in the season and the amount of solar radiation reaching Earth (Chappell and Grindrod, 1983). The general decrease in humidity and these dramatic and relatively rapid climate shifts during the Quaternary have had a significant impact on regolith and landscape. There is a wide diversity of Quaternary sediments, ranging from marine carbonates on the continental shelf, colluvial, alluvial and evaporitic deposits that have accumulated onshore due to the largely endorheic drainage, glacial sediments, especially in the hilly landscapes of the southeast, and vast inland areas of desert sands and other aeolian materials. Quaternary sand dunes and sheets cover some 40% of the continent, and saline playa sediments are common into major Quaternary depocentres. The continent was, however, tectonically quiescent by the Quaternary and volcanic activity was confined to northern Queensland and parts of Victoria and adjacent South Australia (Ollier, 1977).

BEDROCK WEATHERING

General

Many landscapes in Australia, particularly on the western plateau (the Precambrian Shield and the intracratonic basins of shelf and platform sedimentation), retain a deeply weathered mantle. Weathering is generally best preserved in regions of low local relief, reflecting long-term tectonic stability in down-faulted situations, and where they are protected by duricrusts or later sediments. Deep weathering has affected most rocks and geological provinces across the continent. The depth of weathering may be as much as 200 m but varies considerably and is controlled by lithology, structural features, landscape position and any overlying sediments at the time of weathering. For example, weathering is very extensive on the Girilambone Group rocks, except for some prominent quartzite units (McQueen, 2004). The Lower Devonian turbidites of the Cobar Basin are moderately to deeply weathered, but the more siliceous units, equivalents outside the basin, are less weathered. Weathering is generally greatest in the greenstone belts, though the intervening granites and much of the sedimentary cover, including the Mesozoic and Tertiary sediments, are also deeply weathered. In the Yilgarn Craton, weathering of granitic rocks is generally <30 m, except in mineralized areas and along shears (Anand and Paine, 2002). Deeper weathering in mineralized areas is due to the high chemical reactivity of sulphides, intense shearing and variation in competence of contiguous rocks. Joints and fractures act as permeable zones and as conduits for weathering fluids to flush away the soluble products of weathering. The extent of weathering and erosion, and hence landscape expression, is also controlled by geological structures, whether pre-existing or caused by tectonic activity during landscape evolution. For example, Li and Vasconcelos (2002) illustrated the influence of tectonic history on weathering in three regions of Queensland. Mt Isa represents a tectonically stable craton, most favourable for the preservation of deeply weathered materials and has the oldest weathering ages, 70 to 0.6 Ma. Charters Towers represents a tectonically active margin, with relatively high erosion rates (up to 12 m Ma^{-1}; Seidl *et al.,* 1996) that prevent the preservation of ancient weathering profiles (17-0.6 Ma in age). Mt Tabor represents a transitional environment between a relatively tectonically quiescent craton and active marginal zones, giving weathering ages ranging from oldest (27.2-6.8 Ma) in the W to youngest (13 Ma-200 Ka) in the E. Similarly, tectonic processes have had a major impact on the denudation and preservation of weathered materials and their subsequent landscape expression in western NSW and the Mt Lofty Ranges in SA (Hill, 2000; Gibson, 1997; Hill, 2005; Tokarev and Gostin, 2005)

In general, weathering is deeper on palaeoplains and topographic lows than on erosional plains and hill belts. In many places in the Mt Isa region, Proterozoic bedrocks are weathered to greater depths where overlying Cambrian or Mesozoic sediments have been removed or were never deposited.

Regolith formation on Precambrian basement

Deep weathering of Archaean and Proterozoic bedrocks results in a regolith comprising a number of zones formed by structural and mineralogical modifications of the bedrock. During weathering, some of the components of primary minerals are leached and secondary minerals are formed as residua (Figure 6). The pathways by which these

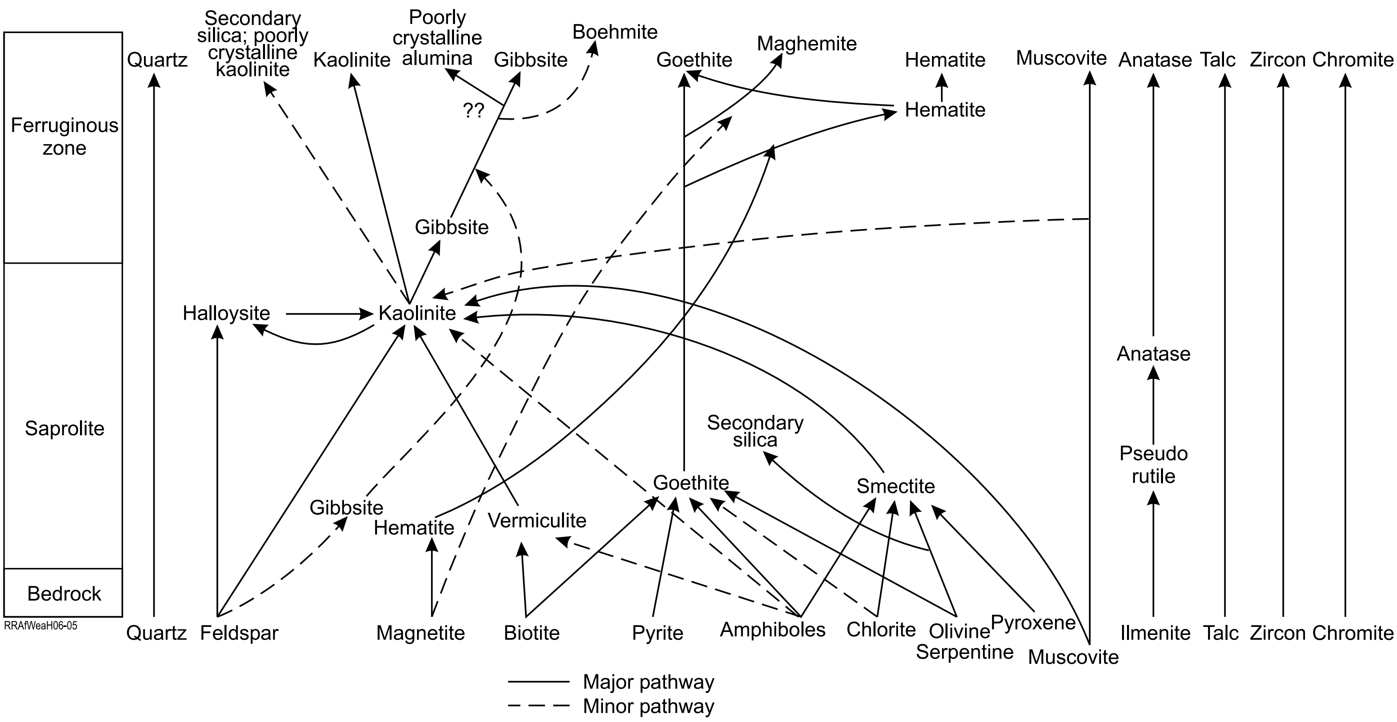

Figure 6. Pathways of formation of secondary minerals in weathering profiles (after Anand and Paine, 2002, compiled from Gilkes et al., 1973; Anand and Gilkes, 1984a, b, c; Anand et al., 1985; Singh and Gilkes, 1991; Robertson and Butt, 1993).

minerals form are varied and complex. The final product of weathering of all rocks is a mineral assemblage of least soluble minerals (kaolinite, hematite, goethite, maghemite, gibbsite, anatase, boehmite) and the most resistant primary minerals (quartz, zircon, chromite, muscovite and talc), although neoformation of several generations of hematite, goethite, kaolinite and gibbsite may occur (Anand and Paine, 2002). Poorly crystalline minerals are an important constituent in surface or near surface materials. The more soluble minerals, including carbonates, sulphates and halides, occur in arid environments. The mineralogy of weathering profiles developed on mafic, ultramafic and felsic rocks is summarized in Figure 7.

Different physical and chemical conditions characterize the various parts of a residual profile. Thus, in the slightly weathered rock at the base of the profile, mineral weathering will take place in microfissures and narrow solution channels, and the capillary water in such spatially restricted volumes may be expected to be close to equilibrium with the primary mineral. In these circumstances, the weathering product formed may be closely related to the primary mineral both compositionally and structurally. In the saprolite, the primary minerals are pseudomorphically replaced by weathering products, retaining the fabric and structure of the parent rock. Clay-rich zones higher in the weathering profile may or may not retain the original fabric and structure of the parent rock but, in either case, the close compositional relationship between primary mineral and weathering product in the slightly weathered rock may be lost. This part of the profile generally will be affected by free-flowing drainage waters in a clay zone, the

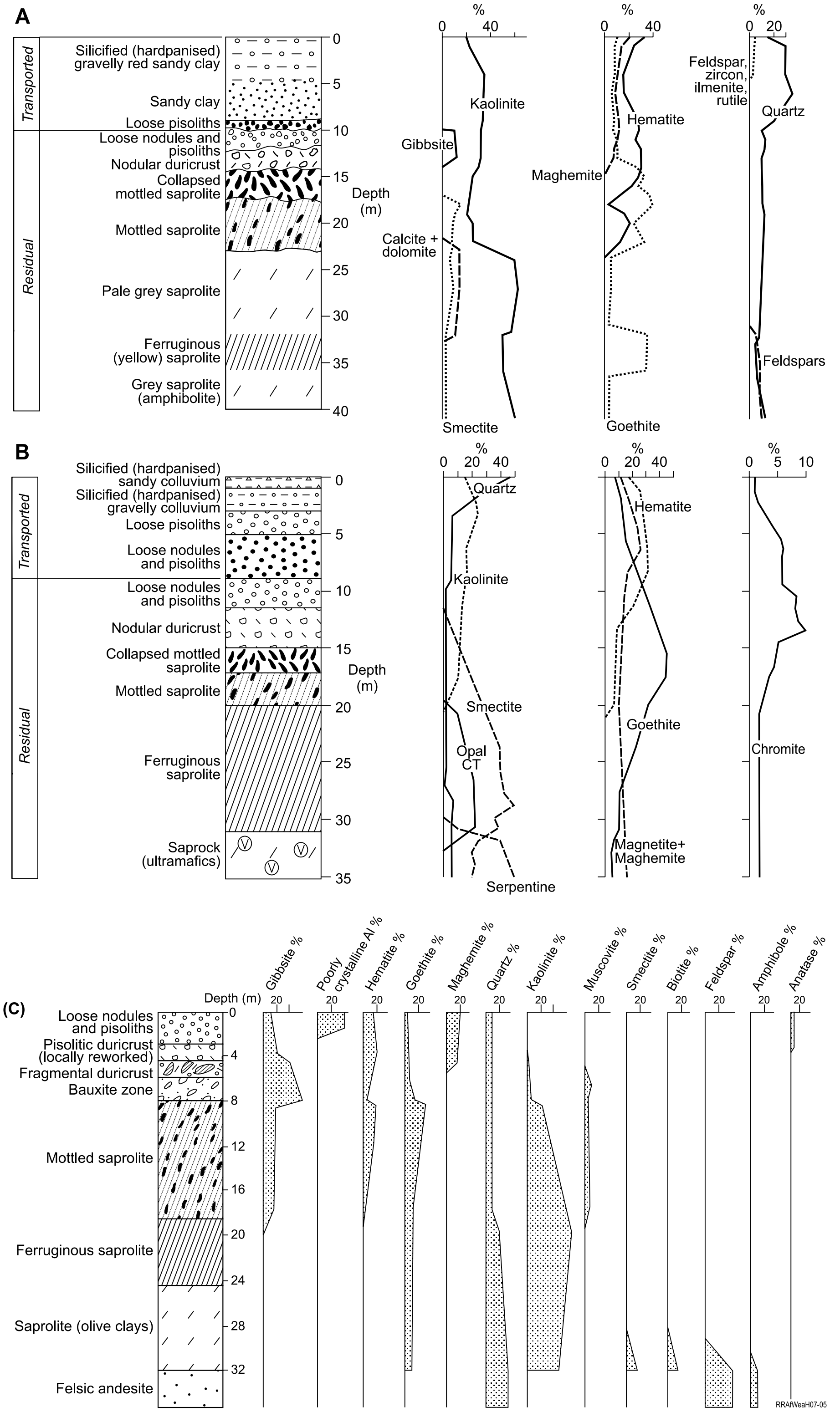

Figure 7. Mineralogy of kaolinitic and bauxitic weathering profiles. A On amphibolite, Lawlers. (B) On ultramafic rock, Lawlers (after Anand et al., 1991). (C) On felsic andesite (after Anand, 1994).

compositions of which will be in disequilibrium with specific primary minerals and the weathering products reflect the interaction between bulk water and bulk parent material. In the ferruginous horizon and soil, the situation will be further complicated by organic ligands derived from decomposing organic matter, or by the direct activities of soil microbes or plant roots. Biological weathering assumes a much greater significance in the upper horizons compared to mainly inorganic processes in the saprolite and saprock. The general nature of any particular weathering profile will reflect the interactions between climate, topography, parent material, soil biota and time. Superimposed upon this complexity, when considering how individual primary minerals break down in detail, will be factors related to the nature of the mineral itself. Particularly important in this respect is the inherent susceptibility of the mineral to weathering, which is related to overall chemical composition and structure, as well as the distribution and density of defects, dislocations and exsolution features that commonly control the progress of the weathering reaction. Mechanisms of weathering of primary minerals are not discussed here and readers should refer to Anand and Gilkes (1984 a, b, c), Robertson and Eggleton (1991), Robertson and Butt (1993), and White and Brantley (1995).

Walther (1915) summarized a combination of observations into a diagram showing a 'lateritic profile' as an iron crust with a mottled zone and a pallid zone overlying fresh Archaean bedrock. Examples can be seen in many deeply weathered terrains, but the profiles vary considerably. Saprolite is common to all weathering profiles but the nature of the upper horizons may differ with lithology and may even have formed from material unrelated to the underlying bedrock. Profiles with ferruginous duricrusts are not common particularly in the Gawler Craton, Girilambone-Cobar region and western NSW. McQueen (2004) noted that well developed weathering profiles show considerable ferruginization in the upper parts and ferruginous mottling and veining in the saprolite in the Girilambone and Cobar region. However, they cannot be described as classicial 'laterite' profiles. This may be due to the relatively low iron content of the common lithologies (siliclastic sediments) and greater erosional modification during profile development (McQueen, 2004). Such is also the case in western NSW (Hill, 2005). Where classical 'lateritic' profiles do occur (*e.g.*, in the Yilgarn Craton, Mt Isa region), they most commonly comprise fresh rock grading upwards into saprock, saprolite, a clay-rich (plasmic) or sand-rich (arenose) zone and a ferruginous upper horizon.

A 'typical' weathering profile consists of two major components, the saprolith and pedolith, distinguished by their fabrics (Figure 8; Anand and Butt, 1988; Eggleton, 2001; Anand and Paine, 2002). The *saprolith* comprises saprock and saprolite that has retained the fabric originally expressed by the arrangement of the primary mineral constituents of the parent material. *Saprock* is a compact, slightly weathered rock of low porosity (Trescases, 1992) with less than 20% of the weatherable minerals altered. Compared to saprock, saprolite has more than 20% of the weatherable materials altered. The saprolite is, thus, generally a product of a nearly iso-volumetric weathering, as also shown by quartz veins and fabrics in corestones retaining their original orientations. The zones above the saprolite comprise the pedolith and are characterized by secondary fabrics. The boundary between saprolith and pedolith is the *pedoplasmation front*. The principal horizons are the plasmic or arenose, mottled zone and lateritic residuum (lateritic duricrust and lateritic gravel).

The plasmic horizon is a mesoscopically homogeneous component of the weathering profile developed on quartz-poor rocks. It is dominated by clay or silty clay that has neither the primary fabrics of saprolite nor the significant development of secondary segregations such as nodules and pisoliths. It is a transitional zone of settling and consolidation produced between saprolite and mottled zone by loss of fabric without significant chemical and mineralogical changes. *The mottled zone* is that part of a weathering profile having macroscopic segregations of subdominant colour different from that of the surrounding matrix. Mottling is not restricted to the 'mottled zone' but may occur in plasmic, arenose and saprolite horizons. It results from either the accumulation of Fe (brown mottles) or the loss of Fe (white mottles) (Anand and Paine, 2002). Brown mottles developed in plasmic clays are marked by accumulation of Fe and Al oxides as spots, blotches and streaks. They result from pedogenic activity and are formed by migration and accumulation of Fe oxides in the kaolinitic matrix or voids. Some of this kaolinite may be secondary, filling voids previously created in the bleached domains. Mottling progressively destroys pre-existing fabrics, although plasmic or arenose micro-fabrics may be preserved in the mottles. Where white mottles have formed, bleaching commonly extends down cracks, fissures, and root channels, the latter accounting for near-vertical cylindrical bodies of bleached material. The whole material was once oxidized and red or brown; bleaching is a later event. Mottling in saprolite occurs without loss of the primary fabric, so that mottled saprolite is part of the saprolith rather than the pedolith.

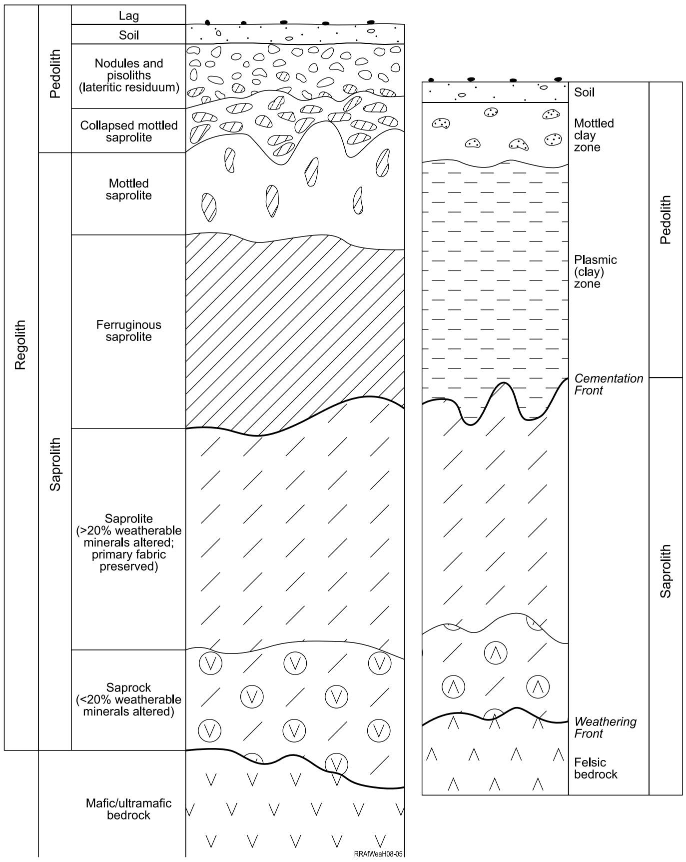

Figure 8. Typical weathering profiles on mafic and felsic bedrocks (modified after Anand and Paine, 2002).

Lateritic residuum is commonly most strongly developed over mafic and ultramafic rocks, but may never have formed on some felsic rocks (Figure 8). Instead, the uppermost horizon is a mottled unit that overlies a white, bleached plasmic clays or an arenose zone. On some mafic rocks, the saprolite more commonly becomes increasingly ferruginous and brecciated, before merging into a lateritic gravel or duricrust. Here, the mostly nodular lateritic residuum has evolved by chemical wasting, partial collapse of mottled saprolite involving local vertical and lateral (generally 5-50 m) movement and the introduction and mixing of exotic materials through soil forming and aeolian processes. Nodules formed from saprolite are commonly lithic (pseudomorphic and lithorelict) and are dominated by hematite, kaolinite and goethite. By contrast, pisoliths formed in soil environments are homogeneous, compound and concentric and have experienced multiple leaching and precipitation of Fe oxides. Pisoliths are hematite and maghemite-rich.

Although much less common, weathering profiles with ferruginous capping have also developed on some Proterozoic bedrocks (*e.g.*, dolomitic siltstone, shale, basalt) in the Mt Isa region, very similar to those on Archaean bedrock in the Yilgarn Craton (Figure 9). Lateritic duricrust is developed on Fe-rich bedrock whereas silcrete has developed on siliceous bedrock (*e.g.*, stromatolitic shale, siltstone), indicating a lithological control. Ferruginous duricrust and silcrete may occur within a few hundreds metres of each other, suggesting that they are coeval (Wilford, 2005; Anand *et al.*, 1997a). The ferruginous profile consist of bedrock, saprock, saprolite, mottled saprolite, and massive or nodular duricrust. Saprolite may be silicified and is largely

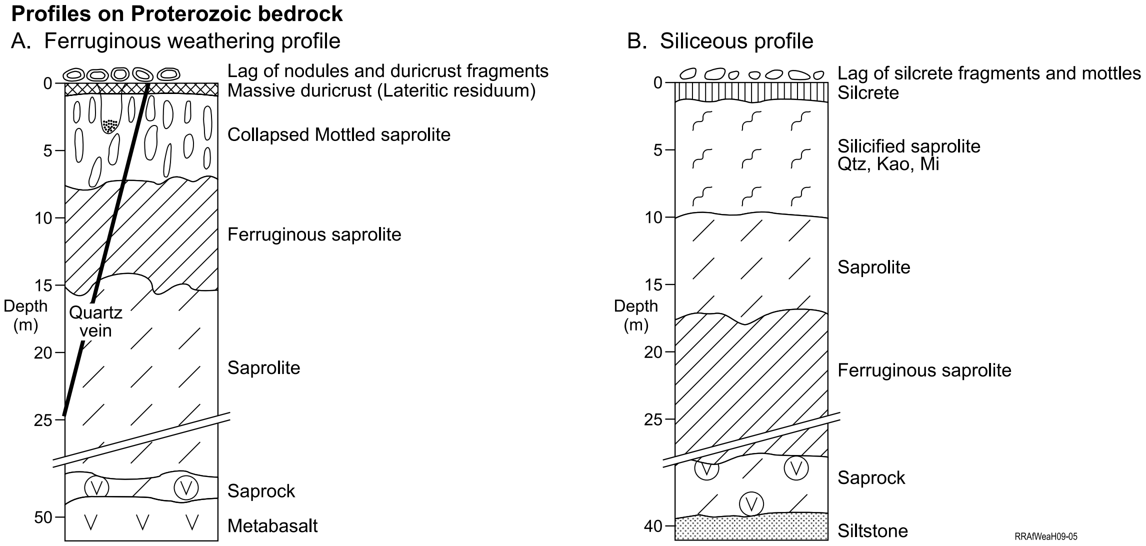

Figure 9. Weathering profiles on Proterozoic basement in the Mt Isa region (after Anand et al., 1997).

kaolinitic but, at the base of saprolite and saprock, smectite and mixed layer minerals occur over particular lithologies (*e.g.*, schists and basalts). There is a gradual transition between mottled saprolite and overlying massive duricrust. Ferruginous duricrusts have formed *in situ* by accumulation of ferruginous materials from mottled saprolite and were let by down wasting of the profile as clays and soluble elements were removed. An idealized silcrete profile consists of bedrock, saprock, saprolite, silicified saprolite, and silcrete (Figure 9). Kaolinitic saprolite is overlain by a bleached quartz-kaolinitic-rich silicified saprolite, with rock fabrics and structures at least partly preserved. Silcrete is an indurated, greyish, massive horizon, rich in quartz and microcrystalline quartz, with small amounts of anatase. Some large massive sheets have columnar jointing. Hematite mottling is common and a lag of mottles and silcrete fragments occur on surface from degradation of silcrete.

CLIMATES AND GEOCHRONOLOGY OF WEATHERING

The distribution of lateritic weathering profiles has been used as a palaeoclimatic indicator. The presence of weathering profiles is generally interpreted to indicate seasonally humid conditions. However, the origin of landforms and regolith in tropical climates is not universally accepted (*e.g.,* Taylor *et al*., 1992; Bird and Chivas, 1993). Deep weathering may occur in cooler climates, but over long periods. It is generally considered that reaction rates double with every 10°C increase in temperature, although more recently, Brady and Carrol (1994) have shown experimentally that weathering rates of silicates may increase two- to five-fold when the temperature increases from 15 - 25°C. Over million of years, these reaction rates would have significant implications on the relative masses of regolith generated under warm-humid versus cool-humid conditions. Precise and accurate geochronological information about the intensity of weathering reactions within the profiles throughout their history may be necessary for determining exact climatic conditions under which these profiles developed (Vasconcelos, 1998).

The ages of deeply weathered profiles are difficult to determine. Typical problems include the lack of suitable minerals and uncertainties regarding closed system assumptions. Despite the above problems, a number of dating methods have been successfully applied to the Australian regolith (Pillans, 2005). Pre-Tertiary ages from oxygen isotopes, palaeomagnetism and stratigraphic dating confirm the great antiquity of regolith in a number of regions. Taylor *et al.,* (1992) described profiles preserved beneath Palaeocene basalts, indicating that weathering dates from the Late Mesozoic on the Monaro; Bird and Chivas (1988) have dated profiles within the highland belt from Late Mesozoic to Late Tertiary, and Young *et al* (1996) dated deep weathering profiles on the S coast of New South Wales as Late Mesozoic. In NW Queensland, the application of K-Ar and $^{40}Ar/^{39}Ar$ dating of supergene Mn oxides indicates that the evolution of weathering profiles in the Mt Isa region spans the entire Tertiary, possibly extending into the Cretaceous (Vasconcelos, 1998). $^{40}Ar/^{39}Ar$ dates of jarosite samples also indicate that weathering-related mineral precipitation continued into the Quaternary. In the Yilgarn Craton, Bird and Chivas (1989) obtained a post-Tertiary age for deep weathering using oxygen isotope analysis of kaolinite, and Dammer *et al.,* (1999) obtained an Oligocene-Holocene age by K-Ar and $^{40}Ar/^{39}Ar$ analysis of K-bearing Mn oxides.

The basic geochronological framework of the Australian regolith is provided by palaeomagnetic dating (Pillans, 2005). This is done by plotting the pole of chemical remnant magnetisms (CRM) of secondary mineral phases on the trace of palaeomagnetic poles of known age - commonly known as the Australian apparent polar wander path (Schmidt and Clark, 2000). The most useful mineral phase is hematite. Pillans (2005) recorded palaeomagnetic determinations of hematite from some 30 sites (mostly mines) throughout Australia, with a focus on the Yilgarn Craton. There is a distinct bimodality in the data, pointing to events around 10 Ma and 50-60 Ma, with some evidence of Mesozoic or even earlier weathering. At around 60 Ma, the area had a cool, wet, temperate climate, with the vegetation dominated by conifer forests and woodlands. The second cluster at Late Tertiary (10 Ma) represents climates that were seasonally drier (though rainfall was still higher than at present) and warmer, with consequent flora changes, as southern Australia drifted to lower latitudes. Thus, the two major episodes of hematite formation in the Tertiary occurred under different bioclimatic regimes. The third episode of hematite formation (<1 Ma) was under a semi-arid to arid climate with *Acacia* and *Eucalyptus* vegetation. This spread of palaeomagnetic ages is revealed at Lancefield gold pit (WA), where saprolite (Archaean basement) is overlain by Permian fluvio-glacial sediments, Tertiary inset-valley clays, and Quaternary alluvium (Anand *et al*., 2004). Similar results have been obtained in the Mt Isa region, by dating of secondary Mn oxides. Vasconcelos (1998) noted three main peaks, centred at the Cretaceous-Tertiary (65-70 Ma), the Late Eocene-Oligocene boundary (35 Ma) and the Early to Middle Eocene (20-13 Ma).

The question as to whether weathering was episodic or continuous has long been controversial. The episodic model implies that the reactions are mostly driven by prevailing water-rock interaction. During times when abundant water is available, weathering is favoured and the mass of newly formed minerals is significant. The continuous weathering model implies that weathering proceeds at a more-or-less constant rate that is independent of climate. Many authors (*e.g.*, Frakes *et al*., 1987; McGowran, 1994; Vasconcelos, 1995; Dammer *et al*., 1999) prefer an episodic model, based on the clustering of dates. Others (*e.g.*, Mabbutt, 1980; Bourman, 1993; Taylor *et al*., 1992) prefer a continuous weathering model, arguing that the episodic distribution of ages simply reflects that only Mn oxides and hematite were dated, and that the low frequency intervals in the plots represent times of precipitation of other minerals (*e.g.,* clay minerals). This argument implies that fundamentally distinct mechanisms control the precipitation of Mn oxides and hematite than the other minerals, and cannot easily be resolved without techniques that can date other minerals, such as

TABLE 1

ELEMENT MOBILITY DURING DEEP WEATHERING UNDER HUMID CONDITIONS

Host minerals	Leached	Partly retained in secondary minerals
Released at weathering front		
Sulphides	As, Cd, Co, Cu, Mo, Ni, Zn, S	As, Cu, Ni, Pb, Sb, Zn (Fe oxides)
Carbonates	Ca, Mg, Mn, Sr	
Released in lower saprolite		
Aluminosilicates	Ca, Cs, K, Na, Rb	Si, Al (kaolinite); Ba (barite)
Ferromagnesians (pyroxene, olivine, amphiboles, chlorite, biotite)	Ca, Mg	Fe, Ni, Co, Cr, Ga, Mn, Ti, V (Fe and Mn oxides)
Released in upper saprolite		
Aluminosilicates (muscovite)	Cs, K, Rb	Si, Al (kaolinite)
Ferromagnesians (chlorite, talc, amphibole)	Mg, Li	Fe, Ni, Co, Cr, Ga, Mn, Ni, Ti, V (Fe oxides)
Smectites	Ca, Mg, Na	Si, Al (kaolinite)
Released in mottled and ferruginous zones		
Aluminosilicates (muscovite, kaolinite)	K, Rb, Cs	Si, Al (kaolinite)
Fe oxides	Trace elements	
Retained in stable minerals (all horizons)	B, Cr, Fe, Hf, K, Nb, Rb, REE, Th, Ti, V, W, Zr	

(after Butt et al, 2000)

kaolinite. They also argue that observed deep profiles are more a function of preservation than episodic weathering. However, an inevitable conclusion is that weathering must be continuous but that the rate and intensity would have changed over time in response to climate and other factors. Thus, in arid periods, the low throughput of water becomes limiting, so that weathering and the formation of new minerals are small to insignificant.

EFFECTS OF WEATHERING ON ELEMENT DISTRIBUTIONS

The behaviour of major and minor elements during lateritic weathering has been described by several workers (*e.g.* Sadleir and Gilkes, 1976; Butt and Nickel, 1981; Davy and El-Ansary, 1986; Anand, 1994; Butt *et al*., 2000). The principal effects of deep lateritic weathering (non-bauxitic) on element distributions are summarized by Butt *et al*., (2000), which relates leaching and retention of a range of elements to mineral transformations in the principal horizons of the regolith profile (Table 1). In compiling this table, they noted that this summary is a gross simplification and emphasized that no mineral is entirely unaffected by weathering, no element is entirely leached from any regolith horizon and no element is entirely immobile. Many of the geochemical characteristics of regolith can be related to the development of the lateritic profile under humid tropical to (possibly) temperate climatic conditions of higher water-tables, whereas others are due to later events related to more arid environments with lower water-tables and may still be active. The features produced by these later events appear as modifications of the pre-existing lateritic profile and tend to be reflected more by the minor components of the regolith.

The main trends include: -

- Sulphides are some of the most unstable minerals in humid, oxidising environments and typically persist in the profile only if preserved within vein quartz or resistant lithorelicts. This is consistent with observations that S has been strongly leached from the deepest levels of the regolith and appears to be the element most susceptible to weathering. Many elements hosted by sulphides (*e.g.*, Cd, Co, Cu, Mo, Ni, Zn) are commonly leached deep in the saprock, although a proportion is retained in Fe oxides derived from the sulphides. Carbonates are similarly highly susceptible to weathering, hence elements dominantly or significantly hosted by them, such as Ca, Mg, Mn and Sr, are strongly leached. Calcium and Sr, for example, are thus commonly reduced to very low concentrations throughout most of the regolith.
- Weathering in the lower saprolite causes destruction of feldspars and ferromagnesian minerals. Sodium, Ca and Sr are leached from the former, with Si and Al retained as kaolinite and halloysite. In addition, K, Rb and Cs will be lost if hosted by orthoclase or biotite but, if present mainly in muscovite, concentrations are maintained or residually increased throughout much of the regolith. Barium, commonly hosted by feldspars, is released early during weathering, but is reprecipitated as barite, which remains stable through the regolith and is only partly leached in the ferruginous duricrust. Weathering of less stable ferromagnesian minerals (pyroxene, olivine, amphibole and biotite) produces Fe oxides with partial retention of minor and trace elements such as Ni, Co, Cu, Mn and Ni, and progressive loss of Mg and Si, except where retained in smectite (Mg, Si), kaolinite (Si) or quartz (Si). Brand (1997) observed that, in ultramafic rocks, the Mg discontinuity marks the top of the Mg-saprolite and is characterized by the loss of Mg-bearing phases (silicates and carbonates), a sharp decline in Mg concentration, and a coincident relative increase in Fe oxides and silica. The 2:1 layer of smectites can include a variety of cations having comparable size to the major octahedrally coordinating cations Mg^{2+}, Al^{3+} and Fe^{3+}. Paquet *et al.,* (1987) detail the involvement of smectites as early hosts for selected trace elements during the development of the weathering profile. Trace elements accommodated in the octahedral sheet include Zn, Mn, Co, Ni and Cu. A similar relationship was also observed by Kelly and Anand (1995) in weathering profiles developed on ultramafic rocks at Forrestania. On the basis of electron microprobe analysis of ferruginous saprolite, they suggested that, initially, Ni is associated with smectite but, with further weathering, the association with Fe oxides increases. Brand (1997) suggested that Ni is hosted by eight different phases in the regolith with Mg silicates, magnesite and Fe oxides being the dominant controls on its distribution.
- The alteration of all but the most resistant primary minerals occurs in mid to upper saprolite zones; in addition, less stable secondary minerals such as smectite are also destroyed. Serpentine, magnetite, ilmenite and chlorite are progressively weathered through the zone, but talc and muscovite may persist through to the mottled clay zone and, in places, to the ferruginous duricrust. Ferromagnesian minerals are the principal hosts for transition metals such as Ni, Co, Cu and Zn in sulphide-poor mafic and ultramafic rocks and are retained higher in the profile than sulphide-hosted metals. Nevertheless, they become leached from the upper horizons and reprecipitate with secondary Fe-Mn oxides in the mid- to lower-saprolite.
- Most remnant major primary minerals, except quartz, are commonly finally destroyed in mottled and ferruginous duricrust zones. These zones, in particular, demonstrate one of the principal characteristics of deeply weathered regolith profiles, namely domination by Si, Al and Fe, resident in kaolinite, quartz, Fe oxides (goethite,

hematite, maghemite) and, in places, gibbsite. The abundances and distributions of Si, Al and Fe oxides broadly reflect their lithologies. The distributions of several minor and trace elements are controlled wholly or in part by the distribution of these major elements, due to substitution or co-precipitation. Thus, Cr, As, Ga, Sc, and V tend to accumulate with Fe oxides; Cr, mainly derived from ferromagnesian minerals, is also associated with neo-formed kaolinite. Many resistate and immobile elements also tend to concentrate with clay and Fe oxides in the ferruginous zones although, for most, no chemical interactions are involved. Thus, the distributions of Cr, K, Hf, Th, Nb, Ta, W, Sn, REE, Ti and V relate wholly or in part to their inertness during weathering, which is due to their relative chemical immobility (*e.g.*, V, Ti) and/or to the stability of primary and/or secondary host minerals (*e.g.*, Zr and Hf in zircon; Ti in rutile and anatase; Cr in chromite; K in muscovite). Their abundances tend to increase upwards through the profile due to the gradual loss of other components, with marked accumulation in lateritic residuum, within which lateral dispersion by colluvial action can occur during the course of profile evolution.

- Characteristics commonly associated with weathering under arid conditions are those related to an excess of evaporation, which results in the accumulation in the regolith of weathering products that would otherwise be leached. Duricrusts, formed by the exposure and irreversible hardening of the ferruginous residuum, or the later precipitation of Fe oxides, silica and carbonates, are important in armouring and protecting the regolith from erosion. Enrichment of Ca, Mg, S and Sr in soils and upper horizons, commonly in the top 2 m, relates to the precipitation of secondary sulphates and carbonates (*i.e,* gypsum, calcite and dolomite). Sulphates may also precipitate in saprolite, as alunite; halite is commonly a trace constituent throughout the regolith. Silica induration of various types is also common in regolith in semi-arid areas. Although geochemical dispersion in arid environments is generally considered to be dominated by mechanical processes, it is evident that, particularly where there is an extensive deep regolith, there is significant chemical (hydromorphic) mobility of many ore-related trace elements in groundwater and soil environments, including U and V (Mann, 1984) and Au.

REGIONAL DISTRIBUTION OF REGOLITH AND LANDSCAPE

Geological provinces

There is a close correlation between the major geological provinces and the nature of contemporary Australian landscapes. The continent can be broadly divided into a western province of predominantly Precambrian rocks, and eastern province of Phanerozoic fold belts and sedimentary basins. Both are overlain by Mesozoic and Cainozoic sedimentary basins containing relatively undeformed flat-lying depositional sequences (Figure 2). The western (Precambrian) province is typified by plateaux and ranges, and the eastern (Phanerozoic) province by the highlands along the eastern seaboard and lowlands on the basins. The Western plateau forms the western two-thirds of the continent and is characterized by extensive plains of low relief, at elevations of 300-600 m. Higher ground, with generally stronger relief, is associated either with distinct structural provinces - for example, the Pilbara Craton and Kimberley Basin or with uplifted margins to the plateau, such as the Darling Range.

In the absence of regionally extensive erosion, deep weathering profiles are largely preserved. Preservation is greatest in the arid interior, beyond the marginal zones of valley incision, on land surfaces of mainly subdued relief. The regolith has broadly similar characteristics throughout the region, being dominantly kaolinitic, and ferruginous in the upper part, with thickness commonly of 20-100 m. Ferruginization (and some associated bauxitization) is particularly prominent in some of the currently more humid regions of the SW and N (*e.g.*, Darling Range, Mitchell Plateau, Pine Creek Inlier) but, in arid regions, ferruginous zones are mainly restricted to more Fe-rich lithologies. Uplands of resistant rocks in the Arunta Block, Musgrave Ranges and the Kimberley Basin have either resisted weathering or any regolith has been removed. High relief in the Hamersley Basin, however, is associated with the preservation of deep, highly ferruginous regolith over Banded Iron Formation.

Phanerozoic rocks of the Eastern Highlands occupy the easternmost 500-600 km of the continent. The highlands are formed of predominantly Lower- and Mid-Palaeozoic, highly to moderately deformed sedimentary and volcanic rocks, intruded by numerous granitoids. The regolith of the highlands is highly variable because of the complex history of erosion, deposition, tectonic and volcanic activity, and the preservation of older land surfaces (Chan, 2005). Many such surfaces have a deeply weathered regolith and have ferruginous and siliceous duricrusts. Some of the ferruginous materials are associated with deep weathering profiles, others not (Ruxton and Taylor, 1982). Silcrete is most commonly associated with sediments below and between basalts flows in the eastern highlands (Taylor and Smith, 1975; Joyce *et al*., 1994). Intra-basaltic weathering profiles also occur in some areas (*e.g.*, Monaro), some mappable over distances of more than 10 km. Most profiles are bauxitic (Taylor *et al.,* 1992). In areas where relief is great, erosion rates are relatively high and the regolith is thin or absent. This is the case where tectonic activity has caused uplift and where river valleys have cut deeply into highlands. Alluvial sediments are preserved on the plateau, for example, in the NW of Bathurst as erosional rises at about 600 m elevation, and on lower plateau remnants (300 m elevation) W of Parkes and Forbes. The sediments are remnants of the Jurassic Surat Basin that covered the Lachlan Fold Belt and infilled a topography more rugged than the present (Gibson and Chan, 1999).

Some of the sedimentary basins flanking the western plateau and eastern highlands have depositional histories dating back to the Palaeozoic, but most owe their present topography to the mid-Cretaceous, which left extensive flat landscapes underlain by detrital sediments. Some basins continued to receive sediment through the Late Cretaceous.

Regolith-landform provinces

The long and stable history of Australia, punctuated by periods of uplift and changes of climate and associated geomorphic processes, has resulted in the widespread development of landscapes of hills, plains and tablelands and a variety of regolith types. In detail, there is great complexity and little order in regolith patterns, but closer study enables the delineation of a number of regolith-landform provinces, each with unique regolith and landscapes. Several processes have operated on each assemblage of bedrocks and sediments to produce a particular landscape and regolith. The regolith materials that make up these provinces were either horizons of a deep profile developed by *in situ* weathering of bedrock, or consist of unconsolidated, transported debris of various ages, parts of which are secondarily cemented.

For convenience, three broad geomorphological subdivisions are recognized, the nature and characteristics of which are shown in Figures 10-15. These are: -

- Hill belts
- Dissected palaeolandsurfaces. These can be subdivided on the basis of their dominant geology into: -
    a) Dissected palaeolandsurfaces on basements (Archaean and Proterozoic)
    b) Dissected palaeolandsurfaces on Mesozoic sediments
    c) Dissected palaeolandsurfaces on Tertiary sediments
- Plains: mantled with sediments of various ages that are underlain by weathered or fresh basement.

The term *landsurface* implies that all points in a given surface were exposed to the same climatic conditions at a particular time. Weathering conditions may have varied at different points on the surface, depending on its topographic position and aspect within the surface. The term does not imply a flat, even, featureless regolith-atmosphere interface.

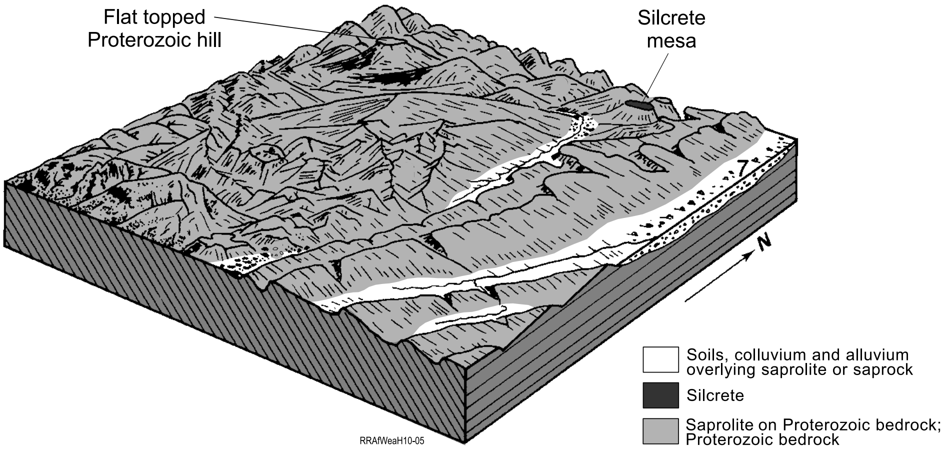

Figure 10. Block diagram showing the nature of the regolith and landscapes in hill belts.

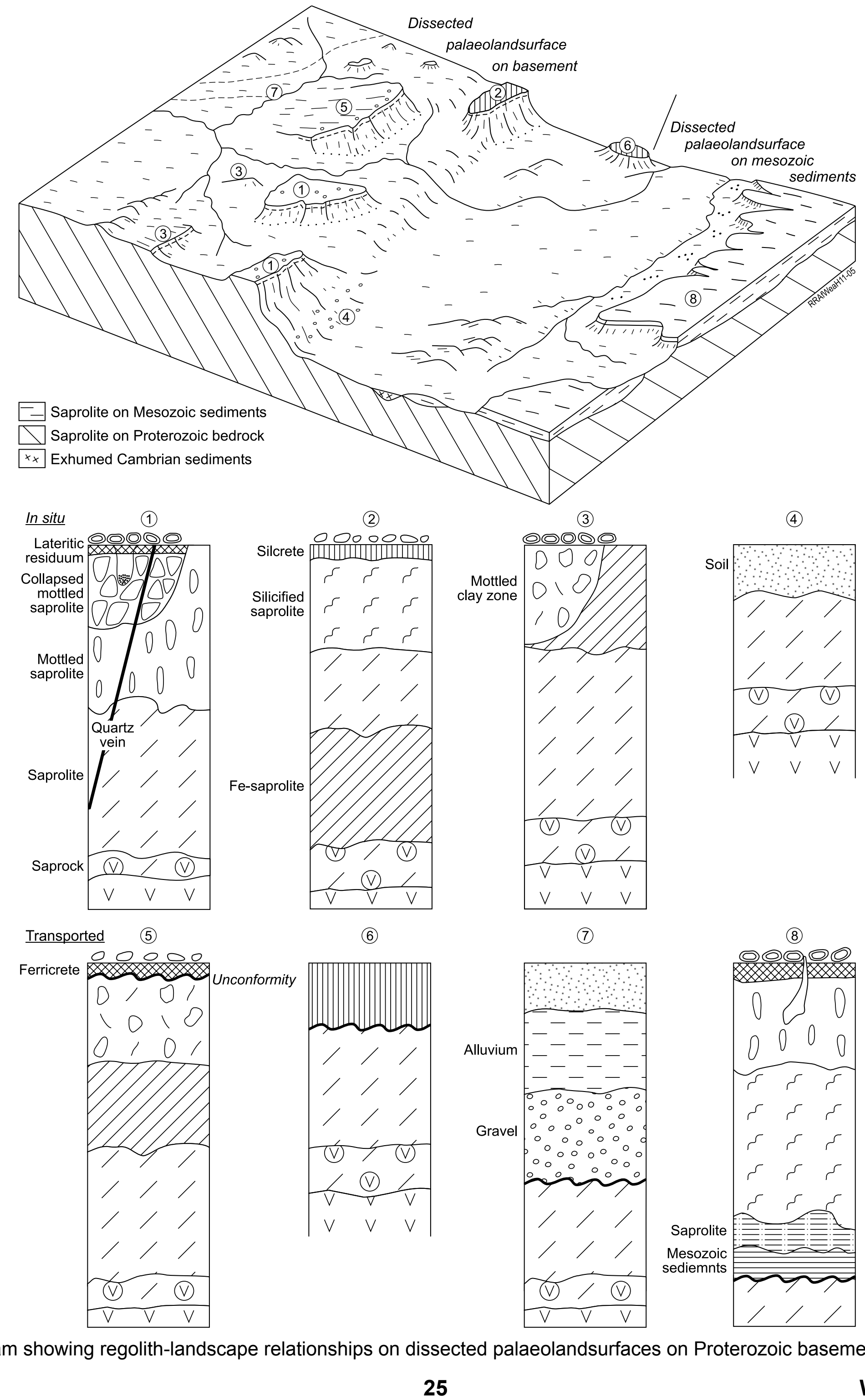

Figure 11. Block diagram showing regolith-landscape relationships on dissected palaeolandsurfaces on Proterozoic basement, Mt Isa region.

Hill belts

Hill belts form deeply incised landscapes consisting of ridges and hills (Figure 10). The detailed sculpture of the landsurface is intimately related to differential weathering and erosion of various rock types. Quartzite, for instance, commonly forms upstanding masses, whereas siltstone and shale tend to offer little resistance and are eroded relatively easily to form lowlands and valleys. In places, some small plateau remnants show a marked concordance of crests. For example, Mesozoic sediments in the Mt Isa region rest unconformably on Proterozoic rocks as small, isolated mesas.

Not all rocks in the hill belts are fresh; many show signs of former and present weathering, including superficial secondary silicification. Lithosols are associated with resistant rocks and areas of high relief and steeper slopes. Ferruginous nodules in the colluvium imply derivation from a deep regolith formed under earlier humid climates, whereas calcite, dolomite and smectite in saprock and soils suggest formation under conditions of impeded drainage and mild leaching, characteristic of a semi-arid climate.

Dissected palaeolandsurfaces on Archaean and Proterozoic bedrock

These have generally low to moderate relief and show signs of the variable extents of former deep weathering. These areas are part of a dissected plateau on which the original surface forms a significant component of the landscape (Figures 11-13). Exposed regolith is associated with erosional landforms such as erosional plains, escarpments, rises and hills. The regolith is mainly mottled saprolite, saprolite and saprock with isolated ferruginous or siliceous duricrust-capped mesas. Mottled saprolite and saprolite have been exposed by erosion of a pre-existing weathered material, or may represent the most weathered form of the parent rock, which had never been capped with a ferruginous duricrust. Incomplete removal of detritus, eroded from the weathered mantle, has left a widespread, generally shallow, sedimentary cover in adjacent colluvial-alluvial plains and valleys.

Erosional scarps form important contacts between different types of regolith, with generally older and typically indurated (Fe or Si) regolith on mesas and younger materials on the pediments below (Figures 11-13). Strongly weathered profiles, with ferruginous or siliceous cappings, occur on stable parts of the landscape (mesas), although the total depth of weathering is variable. By contrast, relatively shallow profiles occur on pediments and erosional plains. The mesas may consist of ferruginous duricrust, indurated Fe saprolite or silcrete over a saprolite. Pediments are covered with ferruginous gravels and much younger, thin residual soils formed from weathering of underlying saprolite. Soils may contain unweathered primary minerals and, in places, fresh bedrock outcrops.

Mottled zone and ferruginous saprolite are commonly exposed on rises (Figures 11-13). Mottled zones have hard, irregular, non-magnetic, hematite-rich, reddish brown mottles, up to 100 mm in length, set in a yellow, kaolinite-quartz matrix. Collapse due to solution of the matrix may have resulted in the mottles becoming increasingly abundant towards the surface, with later fracturing resulting in a nodular lag at the surface. The mid-upper saprolite of ultramafic rocks, particularly of serpentinized dunites (adcumulates), is commonly strongly silicified. Once exposed, the silicified saprolite is resistant to further weathering and erosion and may form erosional landscapes and local topographic inversion.

Where the bedrock is rich in Fe, pseudomorphic replacement of kaolinite by Fe oxides has led to strongly indurated goethite- and hematite-rich ferruginous saprolite. Such cappings are common over Fe-rich shales, basalts and gabbros and may occur on the margins of the uplands. In places, soft clay-rich masses in the ferruginous saprolite have dissolved, leaving numerous voids. The voids weaken the saprolite structure, leading to collapse (collapsed ferruginous saprolite). Further fragmentation results in a surface lag of nodules.

Formation of ferruginous duricrust and silcrete has been complex (Figures 11-13). They commonly occupy similar topographic positions but may have a different origin. In many location in several regions (*e.g.*, western NSW, Gawler Craton, Girilambone-Cobar region, Dundas Tableland), duricrusts are developed in alluvial sediments. In the Yilgarn Craton, ferruginous duricrusts are residual (lateritic residuum) but others are ferruginized sediments (ferricrete). Lateritic residuum (massive and nodular duricrusts; nodular gravel) has formed by accumulation of ferruginous material from the mottled saprolite by down-wasting due to the loss of matrix minerals. The nodules lack complex multiple cutans, a simplicity interpreted to reflect a short weathering history and little to no transportation. Many ferricretes (whether slabby, conglomeratic or pisolitic) represent strongly ferruginized deposits in upper tributary inset-valleys. They are now exposed as isolated hills or as meandering ridges or mesas due to relief inversion. This is largely due to the preferential weathering of mafic and, especially, ultramafic rocks, to form depressions and valleys. These lithologies develop highly ferruginous lateritic residuum which is augmented by the later deposition of detrital gravels and hydromorphically derived Fe oxides to form ferricrete (see Figure 19).

Ferricretes may have a variety of ferruginous clasts derived from the erosion of pre-existing ferruginous duricrust and ferruginous saprolite. They consist of a dark brown to reddish black, hematite-maghemite-rich nodular or pisolitic, clast-supported masses in a dark brown, detrital goethite-rich matrix (Anand and Paine, 2002). Ferricretes are characterized by high mean concentrations of Fe (mean 76.9% Fe_2O_3) and low concentrations of Si (6.5% SiO_2) and Al (4.9% Al_2O_3) (Anand, 1998). These ferricretes typically show high mean concentrations of Ti (5.7% TiO_2), V (1800 ppm) and Cr (4500 ppm); Ti occurs as anatase, rutile and ilmenite, Cr as chromite and in Fe oxides. Some Ti-rich ooliths (2-40 µm), with goethite coatings, occur in the matrix, apparently having crystallized from solution, suggesting mobility of Ti. These ferricretes are Ti and Cr-rich possibly because of sorting during transport.

Silcretes have also formed in residual and transported materials but more commonly in transported materials. Residual silcretes may have floating quartz grains in a matrix of cryptocrystalline silica and anatase, formed from secondarily mobilized Ti. Over granitic rocks, they may also contain significant concentrations of relict zircons and aluminosilicates in the cement (Butt, 1985). Other silcretes consist of quartz pebbles, gravels and sand that suggest that the original material was either a river channel or sheetwash deposit that has been silicified. Such is the case in the many locations at Mt Isa and Cobar regions and western NSW (McQueen, 2004; Hill, 2005). When these sediments have been strongly silicified, they become very resistant to erosion which can result in relief inversion in dissected terrain. Some such sediments contain silcrete clasts, which implies either there were at least two periods of strong silicification, or that there was a protracted period of silicification during which partial erosion, transportation, deposition and silicification occurred.

Ferruginization of the silcrete is widespread and ranges from minor mottling to Fe oxide-cementation of siliceous nodules. Vasconcelos (1998) observed that pediments to the silcrete-capped mesas in the Tick Hill area in the Mt Isa region are draped by strongly ferruginized silcrete breccias (Figure 11). These apparent palaeopediments may suggest a past large-scale erosion of the silcrete mesas, possibly due to increasing relief due to tectonic reactivation of regional faults. If silicification and ferruginization are evidence for arid and humid conditions, respectively, then a humid climatic period must have post-dated silcrete formation. The age of Mn oxides replacing silcretes poses numerical constraints on the minimum age of silcrete formation in the area. The silcretes of the Tick Hill area are at least 38 Ma old, indicating that they are not associated with the postulated transition towards arid conditions in the Miocene. If they do represent arid conditions, they must have prevailed in the Palaeocene or Mesozoic.

The nature of the lag and soil reflect the parent material and the processes that have moulded the landscape. Lag is the residual accumulation of coarse, generally hard, fragments due to the partial dismantling of the upper horizons of the regolith by solution, sheetwash or wind, supplemented by particles moving upwards, *e.g.*, by churning. On mesas and rises, the surface is dominated by coarse (20-50 mm), irregular fragments of mottled saprolite or hardened mottles with

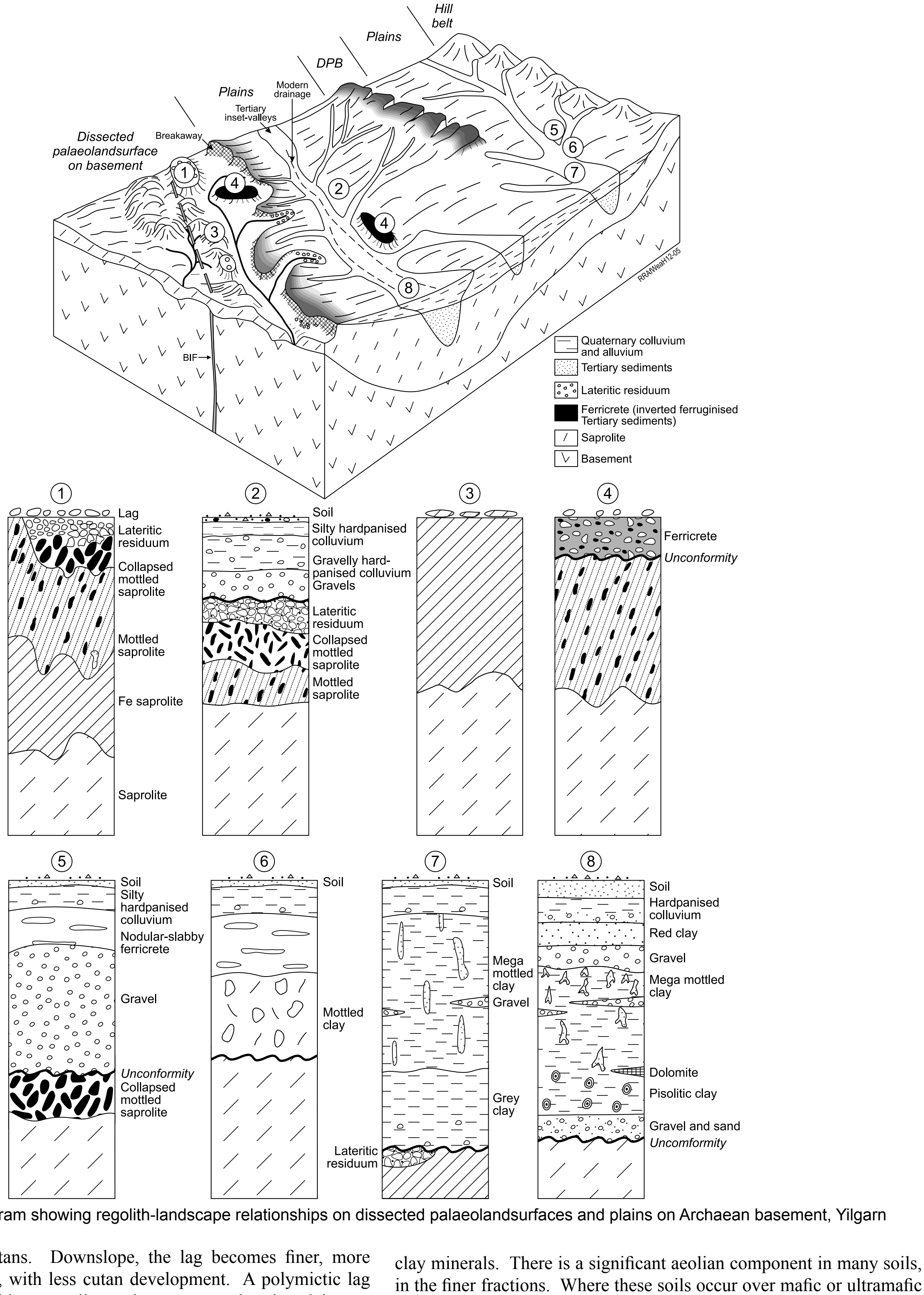

Figure 12. Block diagram showing regolith-landscape relationships on dissected palaeolandsurfaces and plains on Archaean basement, Yilgarn Craton.

yellowish brown cutans. Downslope, the lag becomes finer, more magnetic and darker, with less cutan development. A polymictic lag of ferruginous pisoliths, saprolite and quartz mantles the plains on the gently sloping drainage floors. Much of the lag is dark brown to black, with a smooth, varnished appearance. The fine lag is generally magnetic, whereas coarse lag is non-magnetic. On slopes below breakaway scarps and on hills, the lag is lithic, derived from outcropping saprolite and saprock.

Thin, gravelly soils are developed on ferruginous duricrust and mottled saprolite on broad crests and mesas. Red, brown and yellow earths form on slopes below the plateau remnants, developed mainly from saprolite, saprock and fresh bedrock exposed by erosion. The relief influences the depth of soil, depending on whether material is being received from upslope or lost downslope. It may also affect the drainage status of the soil profile, which may, in turn, determine the nature of Fe oxides and clay minerals. There is a significant aeolian component in many soils, in the finer fractions. Where these soils occur over mafic or ultramafic rocks, the Ti/Zr ratio and quartz contents can be used to identify any aeolian contribution (Scott, 2005).

Dissected palaeolandsurfaces on Mesozoic sediments

The Mesozoic sediments of the inland basins are mostly fluvial and shallow marine, and were deposited in the low-lying areas of the pre-existing landscape. However, in some terrains (e.g, Mt Isa and north-western NSW, Gibson, 2005), partly eroded, weathered sediments are commonly found overlying weathered or fresh basement, forming uplands comprising mesaform hills and interspersed plains with some rocky ridges (Figures 14 and 15; Hill, 2005; Gibson, 2005; Anand *et al.*, 2005; Wilford, 2005; Phang *et al.*, 2005). Such landscapes imply differential erosion, possibly with relief inversion. Because these

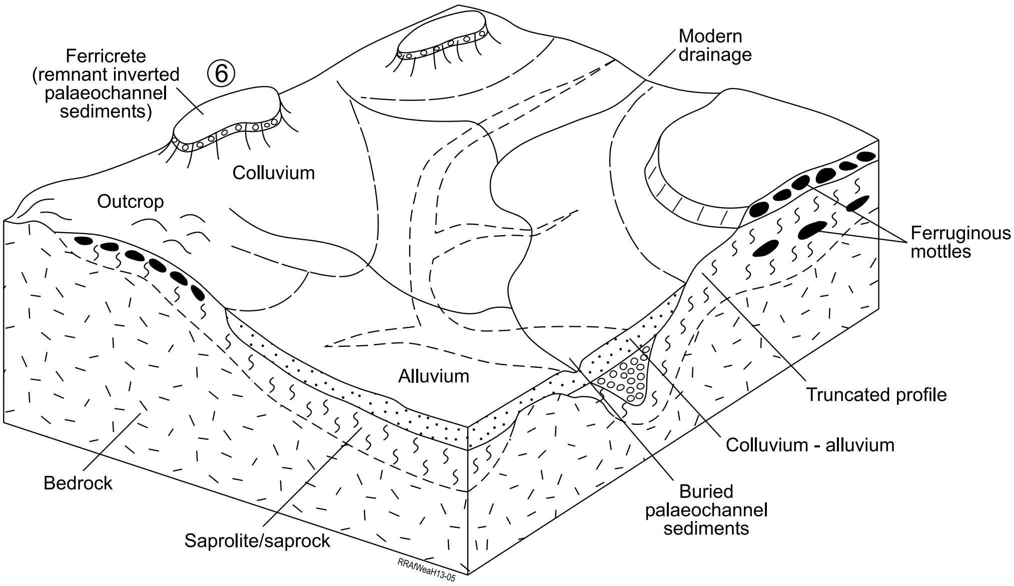

Figure 13. Block diagram showing regolith-landscape relationships on dissected palaeolandsurfaces on basement in Girilambone-Cobar region (after McQueen, unpublished data).

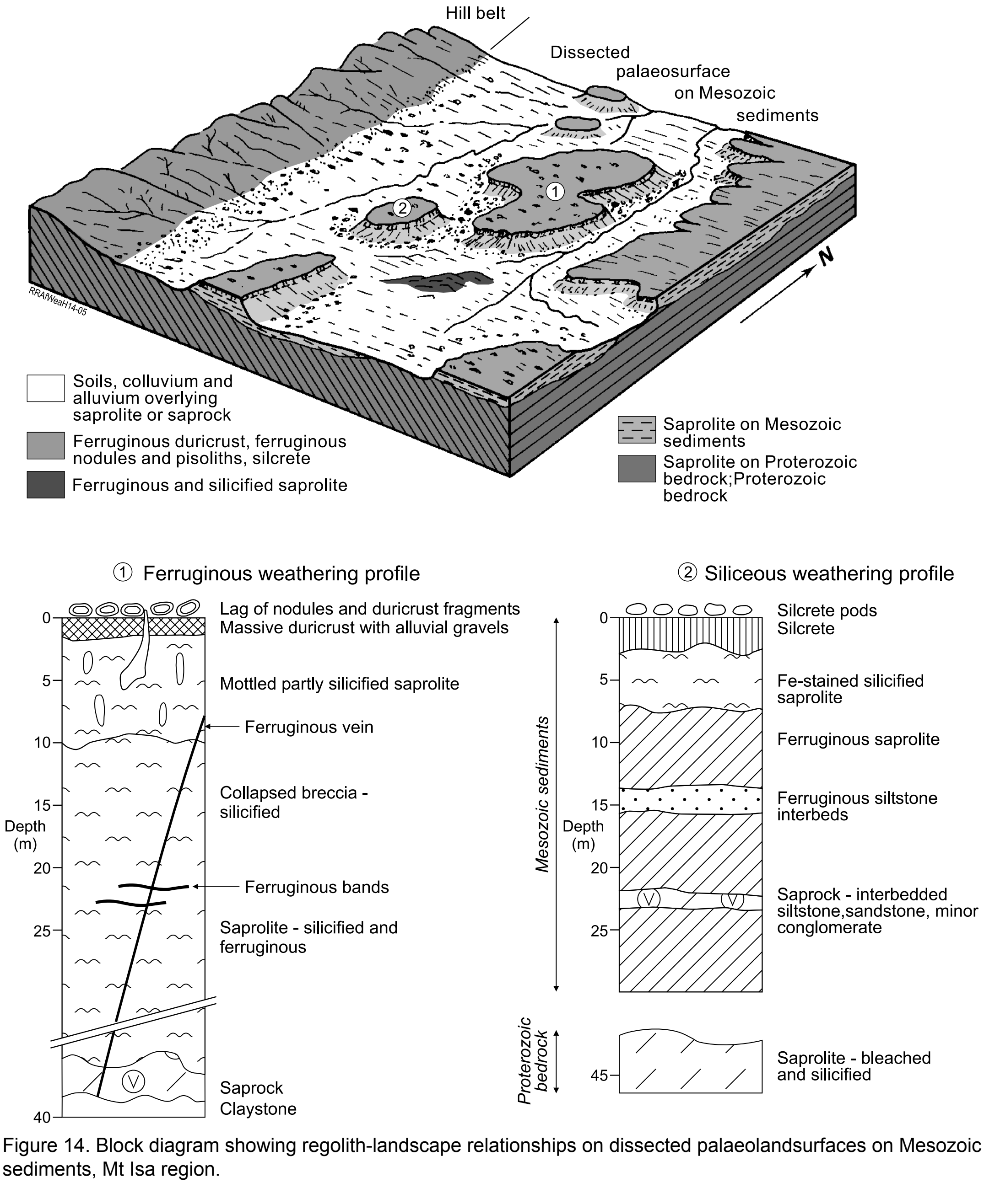

Figure 14. Block diagram showing regolith-landscape relationships on dissected palaeolandsurfaces on Mesozoic sediments, Mt Isa region.

sediments generally cap resistant basement lithologies and have been preferentially removed from areas where they were overlying more erodable and weatherable lithologies, it is probable that they had a much larger aerial distribution on a more subdued palaeosurface than at present (Gibson, Wilford; Anand and others, 2005).

Such landscapes are typical in the Mt Isa region, northwestern NSW and Granites-Tanami region, where mesas and buttes of Mesozoic sediments unconformably overlie basement rocks. These sediment retain evidence for prolonged deep weathering; the upper ferruginized and silicified saprolite contains a well-developed ferruginous duricrust or silcrete (Figure 14). These resistant regolith types are reflected by the presence of local steep low peaks, scarps and breakaways. In the Wonnaminta area in western NSW, most areas of Mesozoic sediments are characterized by a mantle of fragments of silicified and ferruginized sediment and rounded clasts (mostly milky quartz, but also detrital clasts of Devonian sandstone) reworked from the sediment (Figure 15; Gibson, 2005). Some scree slopes beneath silcrete outcrops have been partly cemented to form a thin hardpan with a red earthy matrix enclosing clasts. The hardpan protects the slope from erosion. However, where it has been breached, the underlying saprolite is easily eroded. In the Mt Isa region, the basal unit of these sediments is a silicified conglomerate, up to 5 m thick, comprising pebbles and subangular rock fragments. There are abundant plants, leaf, bark and stick casts on the upper surface of the conglomerate. Ferruginized or silicified Mesozoic sediments occur at several levels. Most are lower than adjacent areas of Proterozoic rocks (Figure 11) but, in some places, they rise to the same elevation. The differences in elevation of the weathered surface suggest that the pre-Mesozoic landsurface had significant relief. The irregular thicknesses of sediments and an undulating Proterozoic-Mesozoic unconformity can be best explained by infilling of valleys. A palaeosurface then formed across the infilled valleys.

Where the Mesozoic sediments have been removed, a veneer of either cherty breccia or silcrete is revealed. This is the base of a Cambrian sequence deposited on silicified Proterozoic saprolite (Figure 11). The eroded Mesozoic sediments now occur on valley floors, ferruginized to goethite and Mn-rich vesicular duricrust.

Weathering profiles on Mesozoic sediments are variable and depend on the sediment type, with thicker profiles developed on claystones and siltstones than sandstones. Ferruginous duricrusts are developed on claystone and sandstone (Figure 14) and closely resemble those developed on Proterozoic rocks (Anand *et al*., 1997a). However, they are relatively enriched in Si due to their high quartz content. The high Fe content of the duricrusts compared to the sediments implies that they have largely resulted from absolute accumulation of Fe and they invariably contain more Fe than would readily have been derived from their Fe-poor parent rocks. The ferruginous profile over claystones contains ferruginous nodules overlying patchy, massive or slabby duricrust that passes down into a mottled saprolite with blocky megamottles. The contact between the hematite-rich duricrust and the mottled saprolite is gradational. Below the mottled saprolite, there is silicified collapsed ferruginous saprolite and the boundary between them is uneven. The collapsed saprolite consists of a siliceous breccia in a yellowish brown clay matrix. This grades down into white, brown or purple saprolite clays. These clays are generally smectitic and contain some kaolinite, goethite and mica. Nearly flat ferruginous bands associated with the saprolitic clays form a steplike microrelief (Phang *et al.*, 2005. The bands consist mainly of goethite, quartz, some mica and kaolinite.

Dissected palaeolandsurfaces on Tertiary sediments

Tertiary sediments occur across much of onshore and offshore Australia, occupying gently down-warped basins and rifted troughs, infilling inset-valleys, and forming widespread, thin sheets in the interior. Where exposed, the sediments are commonly highly weathered, silicified and ferruginized. This situation is typified by the Charters Towers - North Drummond Basin. The Southern Cross Formation (SCF) forms an extensive blanket over the northern Drummond Basin and the Lolworth-Ravenswood Block (Aspandiar, 2005) and was probably deposited during the Early to Mid Tertiary (Grimes, 1980; Henderson and Nind, 1994). It has been extensively dissected, resulting in undulating landsurface with mesas. It is partly covered by the Campaspe Formation. The regolith on the SCF is variable. For example, at Pajingo (*this volume*), it consists dominantly of massive, kaolinitic clays and shales, with cobbles of weathered volcanic material and vein quartz. Cross-bedding of cobble-rich horizons occurs towards the colluvium in the Scott Pit, but it seems that any fine sedimentary features have been obliterated by weathering (Robertson, 1997). The profile has been overprinted by goethite and hematite.

In some places, the SCF has been partly eroded, exposing underlying mottled sediments and mottled basement on breakaways but, in other areas, deep red or yellow earths are underlain by mottled sediments (Aspandiar, 2005). Pisolitic ferricrete is limited to the local high areas on the plateaux and mesas, where incision of the plateaux has increased drainage. Ferruginization occurs throughout the upper part of the profile, with pisoliths present in the top half metre. Some pisoliths are detrital, being derived from older profiles, whereas others have formed *in situ,* similar to those observed for Tertiary sediments in the Yilgarn Craton. In places, for example at Featherby walls (Anand, *et al*., 1997a), the top of the megamottled zone forms a pock-marked surface due to removal of the soft clay material between the more Fe-rich areas.

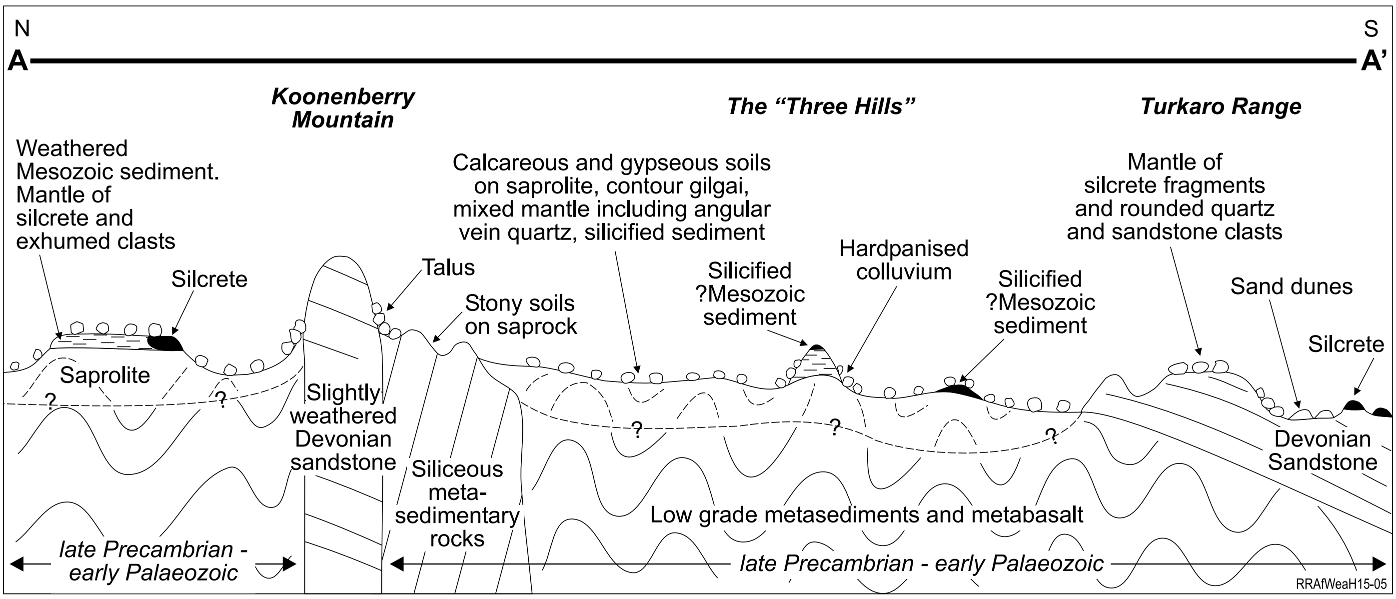

Figure 15. Schematic regolith-landform relationships in N-S zone across the Wonnaminta 1;100 000 sheet in northwestern NSW (after Gibson, 2005).

The Drummond Basin contains a thick sequence of unlithified Tertiary sediments with lignite lenses, the Suttor Formation (SF), which is interpreted to be equivalent to the SCF in the Charters Tower area. It forms discontinuous tracts of isolated mesas and low rises on alluvial plains, or infill valleys beneath younger alluvial material (Figure 16). It can reach a maximum thickness of 120 m (Day, 1981). The SF has been intensely weathered. Ferruginization and silicification occur throughout the whole profile, with duricrusts developed at the top. At Black Creek in the Mt Coolon area, a fine to coarse bedded quartz sandstone with fluvial gravels forms a bluff, capped by duricrust, about 7 m high in which the SF is highly ferruginized with megamottles (Li Shu, 1997). Some of the fluvial sediments have been bleached to white kaolinitic clay. Where Si or Fe has been moved hydromorphically, the bleached sediments have been converted to silcrete (Li Shu, 1997). Field evidence and petrography show that some silcrete has subsequently been desilicified. The complicated weathering processes obscure the original sedimentary structures and make field mapping difficult. Weathering of diverse parent rocks, such as Devonian sandstones or granites, results very similar products.

Plains

General features
This geomorphic province represents low-lying areas with minimum dissection and whose upper regolith may comprise Mesozoic, Tertiary and Quaternary deposits, commonly many metres thick. They may be underlain by regolith developed on basement rocks. The sediments may be highly variable in genesis, provenance, composition and thickness. They may have been derived from erosion of either fresh and weathered bedrock, or by reworking of younger sediments, either locally or from many kilometres away. The sediments themselves may have been subjected to extreme post-depositional weathering. Although the surface distribution indicates a general relationship between the surface regolith types and landforms, evidence from drilling and mine pits in depositional areas show features not evident on the surface (Figure 17). For example, the 3D distribution of regolith in depositional areas has indicated an undulating palaeotopography and a more variable substrate than the present surface would otherwise suggest.

Tertiary and Quaternary sediments are common features of many depositional landscapes of Australia. The following examples are from the Yilgran Craton, Charters Towers – North Drummond Basin, Mt Isa and Gawler Craton regions.

Tertiary sediments: Yilgarn Craton
Inset-valleys (palaeochannels) have long been recognized as a feature of regolith in the Yilgarn Craton. An extensive network of inset-valley, filled with a distinctive sequence of Tertiary marine and non-marine sediments, lies beneath Quaternary cover (*e.g.*, Eastern Yilgarn, Figure 18). In places, inset-valleys overlie, or occur adjacent to, mineral deposits. The relationship between the sediments in the inset-valleys and the deeply weathered regolith developed on basement rocks is important in determining the weathering history of the region. Most of the inset-valleys appear to be steeply incised into deeply weathered surfaces. Apart from being weathered in their upper part, the inset-valley sediments are excellently preserved and constitute the most comprehensive record of Cainozoic depositional and weathering events in the region. The formal stratigraphy and depositional environments of these units have been described by several workers (*e.g.*, Kern and Commander, 1993; Clarke, 1993; Anand and Paine, 2002; de Broekert, 2005; Johnson and McQueen, 2005). The Lefroy and Cowan inset-valleys are characterized by marine sediments and widespread reduced, commonly lignitic units. The sediments of the Roe inset-valley are non-marine.

The Lefroy and Cowan inset-valley systems contain up to 100 m of sediments, the stratigraphy of which is described by Clarke (1993). The Lefroy inset-valley drained to the E into the Eucla Basin, whereas the Cowan inset-valley drained S to the Bremer Basin. The sediments preserve evidence of two marine incursions, thought to have occurred in the Middle and Late Eocene. The Middle Eocene incursion, which is correlated with the Tortachilla Transgression, resulted in the deposition of the Hampton Sandstone/Pidinga Formation and the Norseman Formation in the both the Lefroy and Cowan systems. The Hampton Sandstone/Pidinga Formation has a coarse and fine-grained lignitic and siliciclastic composition and was deposited in a transgressive sequence of fluvial to estuarine sediments, respectively. The Norseman

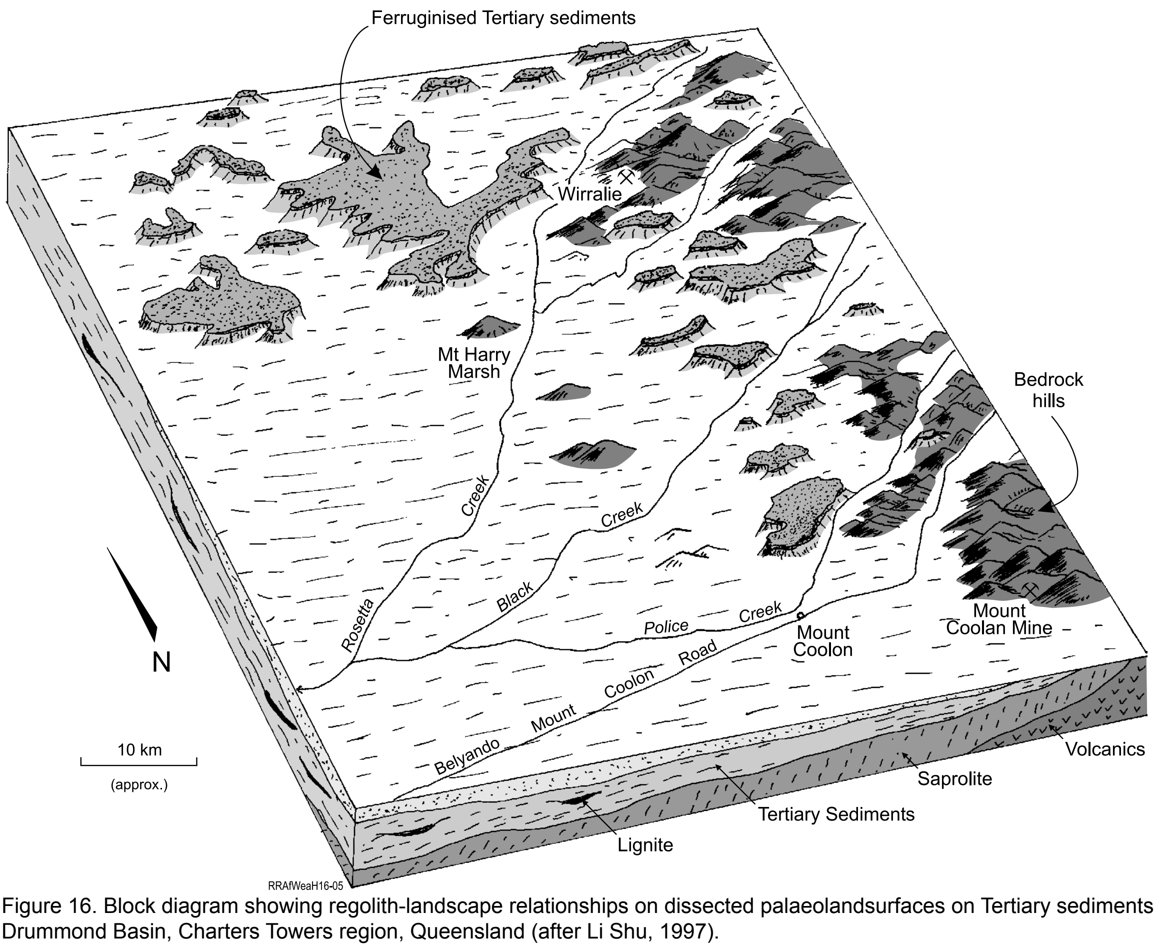

Figure 16. Block diagram showing regolith-landscape relationships on dissected palaeolandsurfaces on Tertiary sediments in the Drummond Basin, Charters Towers region, Queensland (after Li Shu, 1997).

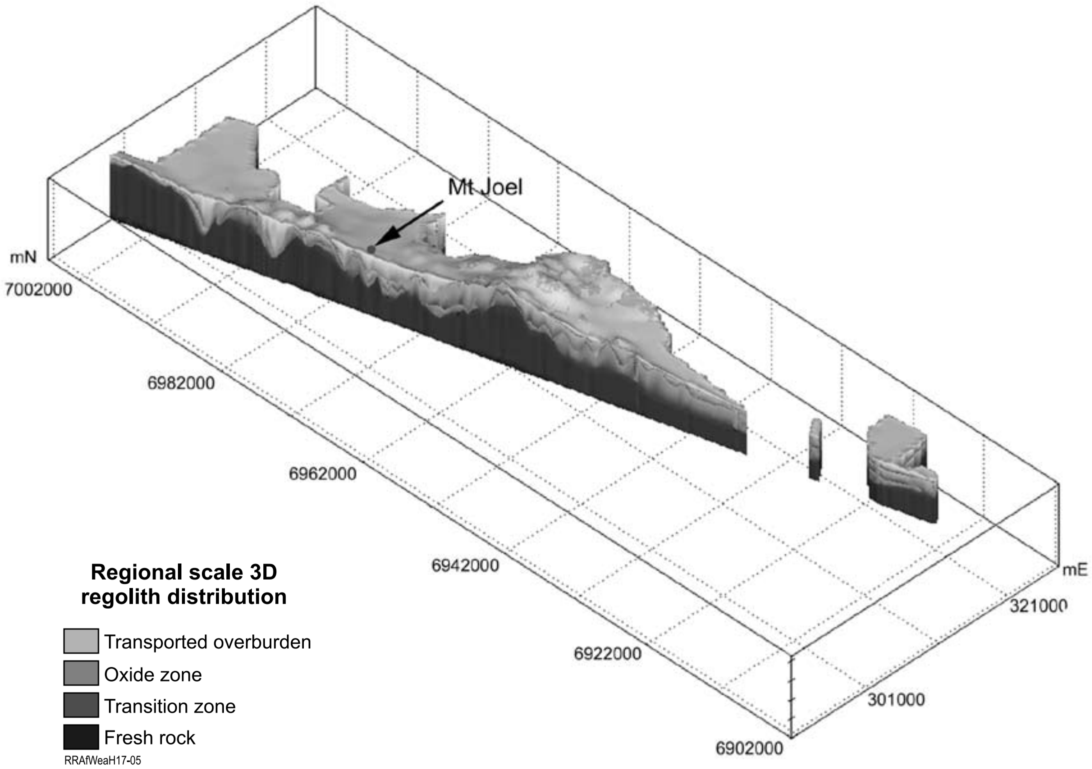

Figure 17. Three dimensional regolith model for part of the Yandal belt, Yilgarn Craton (Anand, 2000).

Formation is a fossiliferous marine carbonate deposited in a shallow, cool water environment (Clarke, 1993). The Late Eocene incursion, correlated with the Tuketja Transgression, resulted in the deposition of the second transgressive sequence. Deposition was initially lignitic, prior to the deposition of the Princess Royal Spongolite

In the Roe inset-valley system, three broad groups of Tertiary non-marine sediments have been recognized (Figure 19) and similar groups and characteristics occur in northern Yilgarn inset-valleys:

Sequence A (sand and gravel). This is equivalent to the Wollubar Sandstone of Kern and Commander, 1993; Alloformation 1 of de Broekert, 2002; Group B of Anand and Paine, 2002). It consists of sand, ferruginous gravel and fragments of ferruginous duricrust, and represents basal fluvial units. It is less common at the base of the trunk inset-valleys than the tributaries. In tributaries, Fe-rich gravels are found over, or a short distance down-valley, of weathered mafic and ultramafic rocks. In places, post-depositional addition of Fe has formed ferricrete (Figure 20) which, when exposed, may occur as low hills or mesas due to relief inversion. Where ferruginous gravels occur in the trunk valley, they are massive to crudely horizontally stratified and may consist of fragments of ferruginous duricrust, ferruginous nodules and vein quartz. These materials may themselves be intensely weathered, with released Fe reprecipitating to form several generations of Fe oxides. Where maghemite-rich gravels are present in ferricrete, they give rise to dendritic patterns visible in aeromagnetic surveys (Anand, 2000). Using current aeromagnetic technology, with a line spacing of 50 m and terrain clearance of 25 m, data from the Yandal belt indicate that a bed of magnetic gravels needs to be more than 1 m thick to show on the aeromagnetic images. In areas of felsic bedrock, inset-valley boundaries are poorly defined magnetically, due to weak or no development of ferruginous gravels, due to paucity of Fe. Erosion of regolith from felsic rocks is likely to be dominated by sandy colluvium (Anand, 2000).

Sequence B (clay to sandy clay with lenses of ferruginous gravel). This is equivalent to the Perkolilli Shale of Kern and Commander, 1993; Alloformation 2 of de Broekert, 2002, Group B of Anand and Paine, 2002). It comprises thick beds of massive kaolinite and smectite, with ferruginous, siliceous and calcareous overprints, and lenses of cross-stratified ferruginous gravel (Figure 19). The clay beds represent a major change in depositional environment from high-energy fluvial to shallow wetland, in which hydrological conditions fluctuated markedly. However, episodes of greatly increased fluvial discharge, presumably related to increase in rainfall, resulted in deposition of the ferruginous gravel lenses (Figure 19). The upward increase in abundance of detrital gravel lenses within clay sequences suggests that the phases of active stream flow became more frequent and long-lived over time (de Broekert, 2002).

A pisolitic clay unit at the base of the Sequence B is widely distributed, although generally absent from the smaller tributaries (Figure 20). It consists of goethitic pisoliths in a grey matrix of kaolinite and smectite with minor amounts of quartz. The pisoliths have high sphericity and typically have well-developed multiple cutans that extend to the centre of the pisolith or encase a nucleus of ferruginous clay, organic debris, quartz, older pisoliths or a mixture of these. Disconformable contacts between sets of cortical laminae, sometimes including thin layers of quartz, indicate that the laminae formed by accretion (Dusci, 1994; Anand and Paine, 2002). This mode of pisolith formation is different to that proposed by Nahon *et al*., (1977), in which cortical laminae develop by the progressive inwards ferruginization of a progenitor particle. Some of the goethite consists of tests of ferruginous bacteria (Anand, unpublished data). Fossilized wood fragments in the goethitic pisolith nuclei indicate the existence of woody vegetation covering the inset-valley filling. Phases of desiccation of wetland facilitated the oxidation of organic matter and growth of goethitic pisoliths. The pisoliths probably formed authigenically within the clay matrix in a hydrologically fluctuating wetland environment.

The kaolinites within the clay unit have a microlaminar structure, suggesting that they are detrital, derived from weathered Archaean basement rocks (Anand and Paine, 2002). In contrast, most of the smectite within the clay unit is probably authigenic, because its delicate lath-like morphology would be unlikely to survive significant transport (de Broekert, 2002). Towards the upper part of the sequence, the kaolinite-smectite clay is highly weathered and intensely bioturbated, resulting in the alteration of smectite to poorly crystalline kaolinite and the formation of megamottles (Figure 19) composed of micro-platy hematite. The mottles are sub-vertical, 0.1 to >1.0 m long, cylindrical in form and in sharp contact with the surrounding kaolinite. Although the clay generally appears to be the matrix in which the ferruginous mottles formed, open-cut exposures indicate that the reverse situation is more likely. The mottling is seen to occur as elongate patches of white clay set in a matrix of red ferruginous clay and to have developed

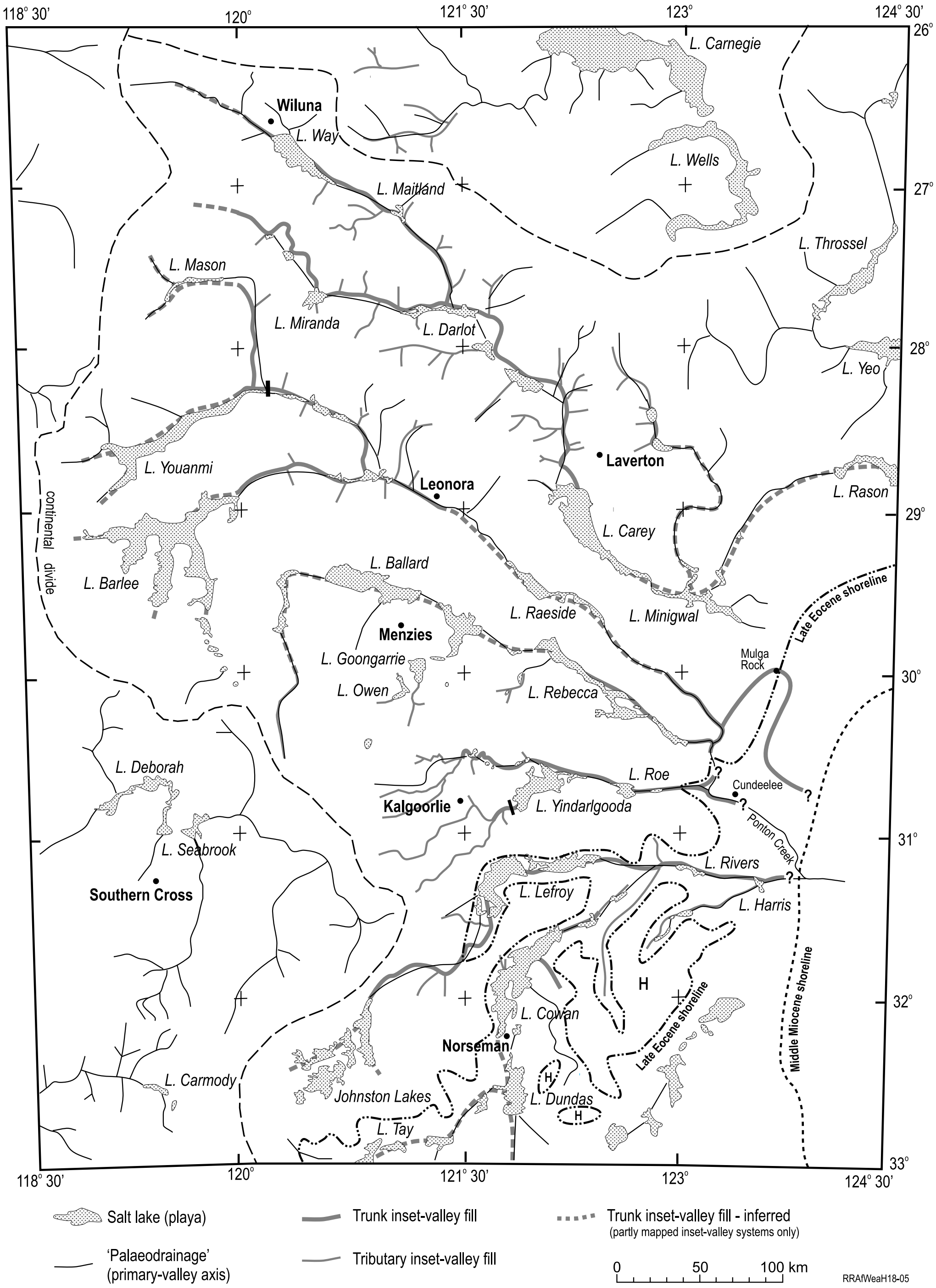

Figure 18. Distribution of Tertiary inset-valley fills in the Eastern Goldfields of Western Australia relative to palaeodrainages, and the late Eocene and Middle Eocene shorelines. Note that only those fills within trunk (about fifth order or higher) inset-valleys are shown (after de Broekert, 2002; compiled from Beard, 1973; Bunting et al., 1974; van de Graff et al., 1977; Kern and Commander, 1993; Johnson et al., 1999).

by removal and reprecipitation of Fe around ancient tree roots (Anand and Paine, 2002). The process was probably affected by the microbial decay of organic matter, which generates reducing conditions under which Fe^{3+} oxides can be reduced, dissolved and redistributed.

Other prominent overprint features of the clay unit may include lenses of nodules of Opal A and Mn oxides, and pods of dolomite. Opal A only occurs where smectite is dominant and is probably formed from silica derived from the dissolution of smectite. Dolomite is precipitated from supersaturated groundwater solutions during the present arid climate, or at least during one of its recent predecessors.

Sequence C (nodular sandy clay). This equivalent to Group C of Anand and Paine (2002). The distribution of these sediments differs from the others in that they are generally exposed at the land surface (Figure 20). Their widespread spatial distribution, sheet-like geometry, crude horizontal bedding and relatively fine grain size, together with evidence of extensive bioturbation, suggest deposition within a low energy fluvial environment, such as the proximal part of a flood plain. They may overlie saprolite with a sharp, highly irregular contact, or unconformably overlie mottled clay with a gradational contact, and, in turn, are commonly overlain by Quaternary sand and gravel (Sequence D, Figure 19). The sediments of Sequence C typically consists of sandy clay to clayey sand, with varying proportion of ferruginous nodules cemented to form nodular-slabby ferricretes. Some nodules are detrital but others are authigenic. Detrital nodules are hematite-maghemite rich whereas authigenic nodules largely consist of hematite, goethite and quartz. Goethite, introduced later, occurs as cutans and may even have entirely replaced hematite nodules. Ferricretes are characterized by higher concentrations of Al (mean 30% Al_2O_3), Si (mean 45% SiO_2) and lower Fe (mean 25% Fe_20_3). They are commonly overprinted by later bleaching. The bleached areas are commonly cylindrical and cored by tubes that may be filled with secondary kaolinite or silica. Widespread preservation of such tubes, presumably fossil root/solution cavity systems, attests to high permeability around the residual and transported regolith boundary. Very fine spherites (Killigrew and Glassford, 1976) of poorly ordered kaolinite, generally 1 mm in diameter, are common in the matrix and nodules of ferricrete and are thought to have formed by concentric deposition of clay particles

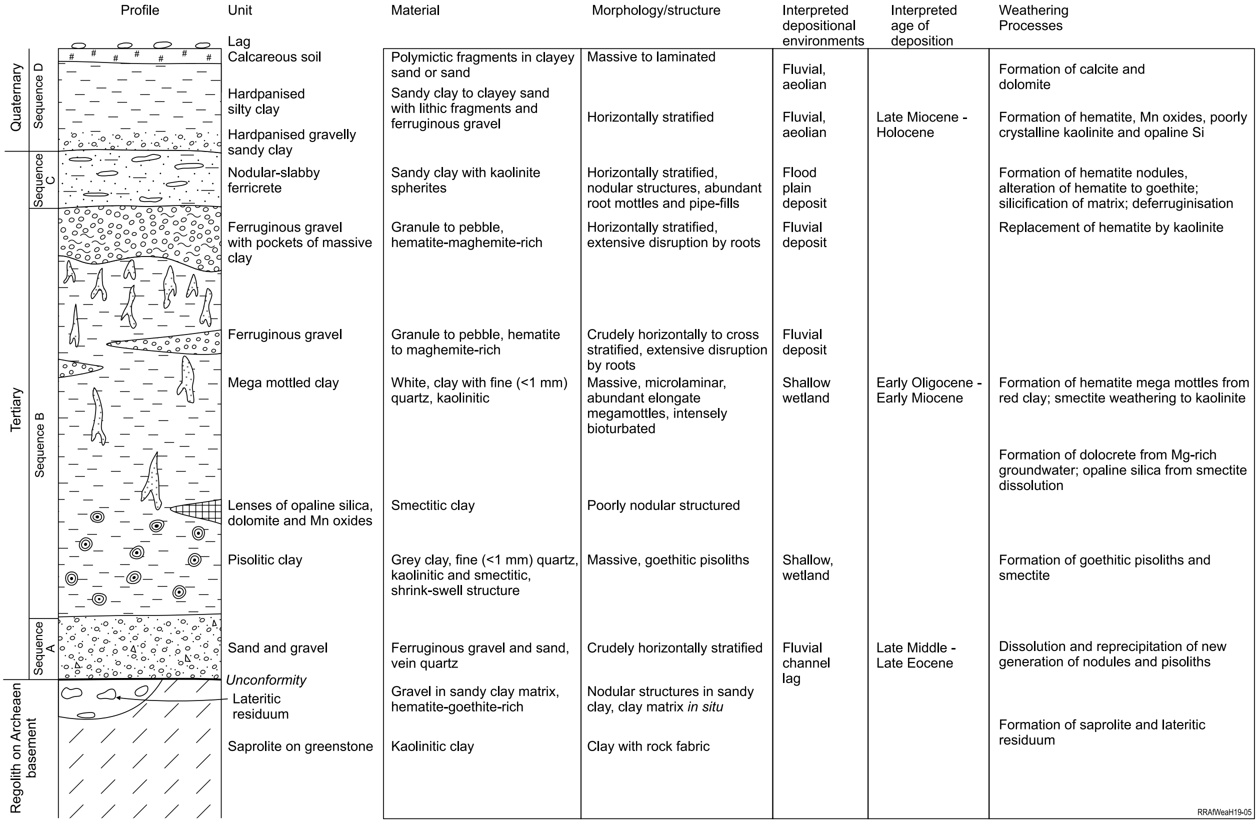

Figure 19. Classification, description and interpretation of stratigraphy of sediments. (modified after Anand and Paine, 2002; Dusci, 1994; de Broekert, 2002).

around a central core of isotropic kaolinite. Killigrew and Glassford (1976) suggested that they are aeolian.

The most conspicuous secondary structures in ferruginized nodular clay are large columns up to 0.75 m in diameter and 2 m long, composed of ferruginous nodule-structured clayey sand, similar to the surrounding sediment (de Broekert, 2002). The origin of these features is uncertain, but similar features elsewhere are interpreted to be pipes produced by the decay of tree roots that have become infilled with detrital sediment. Differential cementation of the pipe fill and a thin zone of the surrounding sediment produces a cylindrical body that is relatively resistant to erosion (Twidale *et al.*, 1999).

Tertiary sediments - Charters Towers-north Drummond Basin

In the Charters Towers area, the Campaspe Formation is an equivalent of Sequence C in the SE Yilgarn Craton. It is dominated by sand, with minor interbedded silt and clay. The Campaspe Formation covers much of the northern Drummond Basin and the Lolworths-Ravenswood Block, and many of the yellow and grey earths on the low plains are related to it (Aspandiar, 2005). The properties suggest dry phases punctuated with short, intense wet periods (Nind, 1988).

The Campaspe Formation has limited outcrop, because it occurs in low, subdued parts of the landscape, with no breakaways. The profile exposed at Red Falls (type section; Wyatt *et al.*, 1970) is one of the few places where it overlies the Southern Cross Formation (SCF). Here, the Campaspe Formation is about 14 m thick and consists of bedded and cross-bedded yellow earths, ferruginous gravels, mottled quartz-rich sediments and sands within a clay-rich matrix. Cementation of quartz grains by Fe oxides occurs at the top of the profile. This ferruginous horizon degrades to a lag of 2-25 mm ferruginous gravels which, in places, are cemented by goethite into nodular duricrust. In contrast, the mottled SCF consists of gritty sandstone and lacks sedimentary structures. A similar profile in the Campaspe Formation occurs at Waterloo and Pajingo (both *this volume*) and in many parts of the Charters Towers-north Drummond Basin (Li Shu, 1997; Anand *et al.*, 1997a).

Tertiary sediments - Mt Isa region

Tertiary sediments are widespread in the plains over the Eastern Succession in the Mt Isa region. In the vicinity of the Eloise deposit (*this volume*), the sediments are strongly mottled, but are not capped

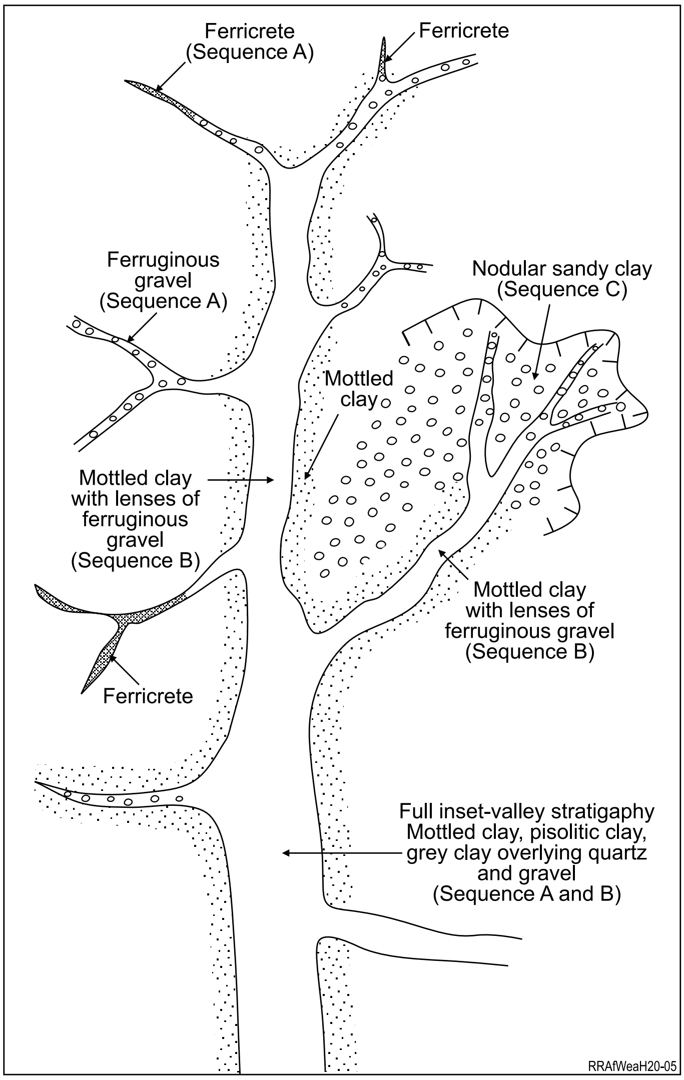

Figure 20. Interpreted relationship between landscape and Tertiary sediments in the Yilgarn Craton. (Anand, unpublished data).

with duricrust (Li Shu and Robertson, 1997). They are less than 5 m thick and consist dominantly of kaolinitic clay, rounded and angular pebbles of rock fragments, ferruginous pisoliths and rounded quartz probably derived in part from breakdown of an earlier lateritic profile. The basal unit is a coarse conglomerate about 0.2-3.0 m thick, consisting of quartz and amphibolite fragments which overlies 40-70 m of weathered, Mesozoic sandstone, mudstone and limestone that become fresh at depth. The thin limestone and many large limestone concretions indicate a shallow water environment. Various fossils in the mudstone and limestone, including belemnites, small bivalves, ichthyosaur remains, shark teeth and ammonities, indicate an early Cretaceous age.

Brown and black soils are common on the Tertiary sediments (Figure 21; Li Shu and Robertson, 1997). Field relationships suggest that both types are found side-by-side across the same sedimentary structure in a similar morphological setting. However, the nature of the soil is a function of its texture that, in turn, determines drainage conditions and the formation of clays. Black soils develop in fine materials and in poor drainage; brown soils develop in coarse materials and well-drained environments. Clay mineral investigations by XRD and SEM confirm that, in brown soil, kaolinite is predominant; in black soils, smectite is significant and kaolinite minor (Anand *et al*., 1997a). It is suggested that, in the initial stages of soil formation, various brown, fine and coarse kaolinite-rich facies of alluvium were deposited on the plains. Where the alluvium was coarse, little water was retained and the alluvium remained brown and kaolinitic; where the alluvium was fine and water was retained, the kaolinitic soil matrix transformed into smectite. Thus, formation of black soil from brown soil is essentially a process of partial transformation of kaolinite to smectite. This interpretation was supported by a detailed SEM examination of theclays. To convert kaolinite ($A1_4Si_4O_{10}$ $(OH)_8$ to smectite ($((Ca,Na)_{0.7}$ $(Al,Mg,Fe)_4$ $(Si,A1)_8O_{20}$ (OH) $_4$ nH_2O), requires addition of Ca, Na, Mg, Fe and Si. Experimental transformation of kaolinite to smectite has been reported by Imasuen *et al.,* (1989). The timing and climate under which this transformation occurred are not known but it is most likely to have been after deposition of Tertiary sediments. According to Imasuen *et al.,* (1989), extrapolation to typical soil temperatures of the tropics would indicate conversion times of one year at 100°C, 15 years at 45°C and 54 years at 25°C.

Gilgai seem to have developed at least in part by differential expansion and contraction of smectitic black soils due to seasonal changes in moisture content, leading to a churning action. Expansion of the soil pushes gravel up the profile; gravel rims to the depressions are common (Paton, 1974; Beckman *et al.,* 1981).

Tertiary sediments-NW Gawler Craton

Tertiary palaeochannels, mostly discharging into the Eucla Basin, are widely distributed in the northwestern Gawler Craton (Benbow *et al*., 1995; Hou *et al*., 2005). The dimensions of the channels vary greatly, with widths of the river valleys ranging from a few tens of metres to more than 30 km and depths of up to 120 m. There are four major units: the Eocene Pidinga Formation (mainly fluvial carbonaceous sediment), Khasta Formation (estuarine sands) and Ooldea sand (coastal barrier sands), and the Miocene Garford Formation (dominantly lacustrine clays). Eocene Pidinga Formation is incised into weathered basement rocks, which are sometimes difficult to separate from the basal Tertiary sediments (Benbow *et al*., 1995). Ferruginization and silicification of Tertiary sediments is common. Ferricrete probably formed in early Mesozoic, late Oligocene-Middle Miocene and Late Miocene-Pleistocene times. Major phases of silicification occurred in late Eocene-Middle Miocene and late Miocene-Pleistocene times, when significant groundwater silcrete formed (Alley *et al*., 1999).

Quaternary sediments and soils

Quaternary sediments (Sequence D, Figure 19) are thinner (0.1-10 m) than the Tertiary sediments, but more heterogeneous and laterally extensive. They may be colluvial, alluvial or aeolian. Colluvial sediments commonly flank hills and rises and extend across the adjacent lowlands. Alluvial sediments are associated with the contemporary drainage network. They may overlie all rock units, whether fresh or weathered, from Precambrian to Tertiary in age, and consist of their physical and chemical weathering products. Quaternary sediments are extensively overprinted by current pedogenic and diagenetic processes. In semi-arid to arid areas, soil development is generally poor, with little horizon development and pedogenesis has been dominated by processes favouring silicification (hardpanization) and carbonate precipitation. In wetter regions, ferruginization is common, particularly in valley floors where dissolved Fe^{2+} derived from surrounding higher areas has precipitated as Fe_2O_3 in the subsoil by percolating groundwaters. The water-table has fluctuated considerably, approaching the surface in the wet months to yield a hydromophic soil. This soil may represent an early stage in the formation of ferricrete.

The development of *hardpans* and *carbonates* occurred late in the weathering history. This is reflected by their presence in residual soils and Quaternary sediment, and by overprinting of ancient regolith materials, such as silicified and ferruginous regolith. Their development is partly related to reduced leaching associated with increasing aridity during the Cainozoic, but also increased input of dust and dissolved components in rainfall.

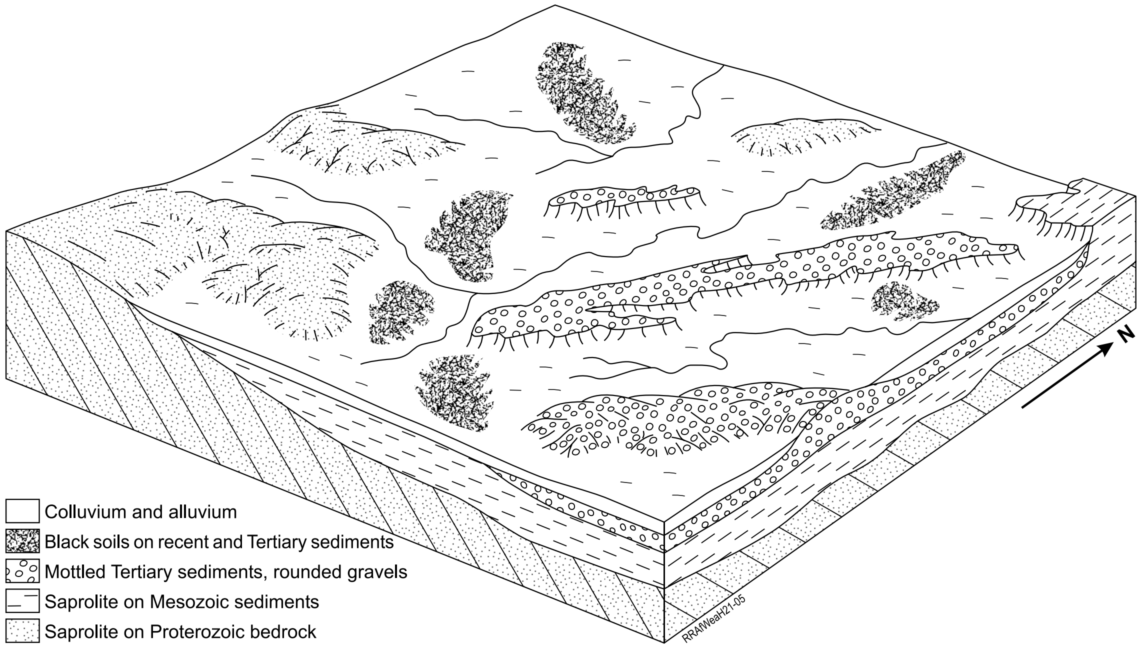

Figure 21. Block diagram showing the nature of the regolith and landscapes on plains, Mt Isa Region.

Red-brown hardpans are common across the northern Yilgarn and the Pilbara, eastward towards the centre of the continent, but are very rare in the south (Bettenay and Churchward, 1974). The southern extent is marked by a gradational boundary referred to as the "Menzies Line" (Butt *et al*, 1977), but red-brown hardpans with calcrete are known as far as 70 km S of Menzies (Mahizhnan and Anand, 2005). This suggests that they were once more extensive and have been subsequently replaced and/or displaced by carbonates. Mahizhnan and Anand (2005) concluded that red-brown hardpan has formed in the Yilgarn Craton where the rainfall is sufficient to dissolve Si and Al, but insufficient to leach them completely from the soil. Dehydration on drying produces a Si-Al-rich gel that precipitates into poorly ordered kaolinite and opal A, similar to that some cemented sands and silcretes (Butt, 1983; Singh *et al*., 1992). Successive dissolution and precipitation leads to the fusion of poorly ordered kaolinites and opal A at the nanometric scale to progressively cement the sediments to a red-brown hardpan. Palaeomagnetic dating of the hematite in red-brown hardpan from the Murchison Province indicates that red-brown hardpan may have been formed prior to the most recent period of normal polarity, from 780 000 years to the present (Mahizhnan and Anand, 2005). However, red-brown hardpan retains its integrity after the removal of Fe oxides (Bettenay and Churchward, 1974), showing that they are not significant as a cement.

Calcrete is extensively distributed in the semi-arid regions of Australia and can be an important sample medium (*see* Sample media – Calcrete and pedogenic carbonate: *this volume*. In most semi-arid and arid regions, the lack of leaching by rainfall leads to substantial accumulations of carbonates in the regolith. Carbonates may accumulate and calcrete form in either the vadose or phreatic zones and vary in nature and abundance, including: i) thin encrustations on freshly weathering rocks, ii) minor diffuse enrichments, iii) friable powders, nodules and pisoliths in soils, iv) sheets of indurated carbonate or carbonate-cemented regolith material. The characteristics and distribution of calcrete in Australia have been reviewed by Chen *et al.,* (2002). Two types are generally recognized, pedogenic (vadose) and groundwater (phreatic) (Sanders, 1974; Butt *et al*., 1977; Arakel, 1982; Anand *et al*., 1997b; Hill, 2000; Chen *et al*., 2002). Groundwater calcretes occur dominantly N of latitude 30°S, particularly in Western Australia, the Northern Territory and northern South Australia, where it forms massive bodies several kilometres long, a few hundred metres wide, and up to 30 m deep, in the axes of major valleys, broadening to form extensive platforms or chemical deltas where they debouch into playas (Butt *et al*., 1977). They form major aquifers in these regions and have significance as sources of potable and irrigation water, and as sites for secondary U mineralization. Precipitation of carbonates at the water-table causes upward heaving, so that the calcrete body forms a positive topographic feature in the valley axis. Calcrete occurrence appears to be related to areas of summer rainfall, where run-off, rapid infiltration and high evaporation limit the period of soil dampness and plant respiration. Dissolved Ca, Mg and bicarbonate in deep groundwater precipitate in drainage axes and depressions after concentration by evaporation or CO_2-degassing due to upwelling or capillary rise (Butt *et al.,* 1977).

Pedogenic calcretes differ from groundwater calcretes in that they may occur in many landform situations. Pedogenic calcretes are especially well developed in semi-arid areas of southern Australia, S of about latitude 30°S, from central NSW and northern Victoria, through South Australia to western Western Australia (Chen *et al*., 2002). They occur near the surface and are laterally extensive for tens of kilometres. In semi-arid regions, the $CaCO_3$ comes from capillary rise and evaporation of $CaCO_3$-charged ground water, from calcareous dust blown by wind and then driven into the soil by episodic rainfall, and from infiltration of soils, sediments, and rocks by runoff from areas containing sources of $CaCO_3$, such as mafic bedrock. In vegetated areas, $CaCO_3$ can precipitate around the roots of plants. The relative contributions to calcrete formation by these processes, and the time relationships represented by the different types of calcrete in general, are not well defined.

Aeolian materials are common in the Australian landscape (*e.g.*, Hill, 2005; Gibson, 2005; Joyce, 2005; Scott, 2005). Widespread dust raising and transport, resulting from high climatic variability and aridity of the Australian continent, are well-documented (Hesse and McTainsh, 2003). Earlier workers believed these sediments to be derived from glacial deposits rather than a desert source. However, studies by Butler (1956) and more recently by Bowler (1976) and Cattle *et al*., (2002) have assigned a desert origin to aeolian material in SE Australia. Approximately 40% of the Australian continent is classified as arid and to accommodate desert landforms, dune fields and playa lakes (Greene *et al*., 2001) that represent major sources of aeolian material. The paths of dust entrainment on the Australian continent and neighbouring oceanic regions are summarized in Figure 22. Dust, clays and silts may be transported over hundreds of kilometres from their original source (Hesse, 1994), whereas sands associated with continental dunes travel much shorter distances and make up only a small portion of the regolith. There is considerable evidence to suggest that much of the aeolian deposition may be associated with individual events. Hesse and McTanish (2003) detail recent occurrences of such events, which are estimated to have deposited millions of tonnes of aeolian material in a single dust storm event. Presumably this material would gradually work into existing soils and, in many circumstances, add significant volume to the existing soil profile.

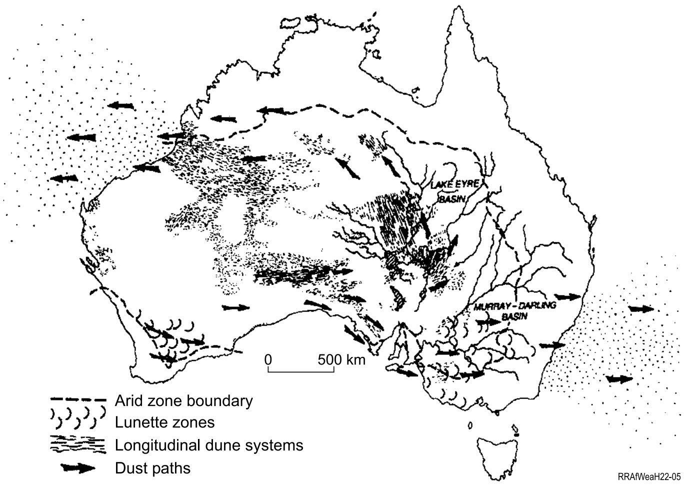

Figure 22. Dust entrainment pathways of the Australian continent (after Hesse and McTainsh, 2003).

LANDSURFACES

Stepped landscapes are a feature of many deeply weathered terrains and their origin has been attributed to repeated periods of scarp retreat and uplift (King, 1953). According to this model, denudation starts at the continental margins, the ultimate base level, and proceeds towards the interior. Episodic uplift and scarp retreat result in a stepped diachronous landscape on which the surface with the highest elevation is the oldest, and those with lower elevations are progressively younger.

Twidale (1956) recognized three major erosion surfaces in the Leichhardt-Gilbert area of NW Queensland. His concept of erosion surfaces was followed by Grimes (1972, 1980) who proposed three major periods of active erosion and deposition for N Queensland, separated by two periods of stability and deep weathering. Although a planation surface developed on a local scale, marked by lateritized Proterozoic and Mesozoic rocks, is clearly visible on aerial photographs, the appropriateness of regional extrapolation and naming of such surfaces is open to doubt (Wilford, 2005; Anand *et al*., 2005; Hill, 2005). In the Buckley River - Lady Loretta area (Western Succession, Mt. Isa region), the landscape is 'stepped' with two main surfaces (Anand *et al*., 2005). Whether these surfaces correlate with those proposed by Grimes is problematical, since erosional surfaces will have different ages in different parts of the landscape. Furthermore, the presence of Mesozoic sediments capping some mesas and plateaux suggests that denudation was not homogeneous. In many places, the pre-Mesozoic landsurface or a derivation similar to it, has been exhumed. In the Selwyn region, several stages of exhumation can be recognized, depending on the relative rates of erosion and the original thickness of the Mesozoic sediments (Wilford*,* 2005).

Hays (1967) argued for the existence of four erosion surfaces throughout the northern half of the Northern Territory - Ashburton, Tennant Creek, Wave Hill and Koolpinyah - representing three episodes of uplift and erosion. An alternative hypothesis of land surface development for the region is proposed by Nott (2005). He suggests that, throughout the Cainozoic, the landscape appears to have increased in relief through gradual etching, not by the multiple episodes of cyclical planation argued by Hays. Nott argues that the type sections for the Wave Hill and Koolpinyah surfaces have been misinterpreted by Hays; instead of being composed of detrital laterite, these type sections are composed of *in situ*, weathered Cretaceous strata, providing convincing evidence against the multiple planation hypothesis. Landscape changes have been driven by the initial deposition and later stripping of marine sediments since the Mesozoic.

In the Hamersley Province of Western Australia, Twidale *et al.,* (1985) postulated the former presence of a high level lateritic "Hamersley 1 Landscape" of subdued relief, and that modern valleys hosting canga and Tertiary pisolitic crusts were incised after removal of this crust (Figure 23). Killick *et al.,* (1996) observed that physical evidence supporting the postulated Hamersley Surface is patchily preserved in the areas they investigated (Figure 23). High level areas have generally been stripped back to relatively fresh rock, although some saddles and other local topographic depressions retain pockets of *in situ* or locally derived weathered material. They reported that weathered zones (not noted by Twidale *et al*., 1985) are best preserved in the base and lower margins of the valleys, and as the ferruginous crusts and underlying saprolite of the pisolitic Tertiary sediments. These *in situ* weathered elements patchily link with the 'smooth' rolling surfaces preserved on some Banded Iron Formation units at high- and intermediate-levels. There is no evidence to support the landscape model proposed by Twidale *et al.,* (1985). In the absence of bedded mineralization, their proposed removal of the laterite zone prior to valley incision leaves no source for mature detritus. The preservation of weathered zones overlain by canga and Tertiary pisolite crusts in the flanks and floors of the valleys, and their conformity with the current topography, indicate that significant elements of the modern Hamersley landscape have been stable since the Tertiary and earlier.

One major difficulty in testing the existence of landsurfaces is the poor age constraints on the formation of the surfaces now exposed (Twidale, 1956). Sedimentary or volcanic deposits on each surface may yield a minimum local age at that point (Summerfield, 1991); however, datable deposits are rare. An alternative method for dating land surfaces is the application of weathering geochronology (Vasconcelos, 1999). Minerals precipitated by weathering reactions may be preserved in the surficial environment and record the minimum age for the exposure of a land surface. Recently, Vasconcelos and Conroy (2003) employed Ar/Ar geochronology of supergene minerals in the Dugald River region of Queensland and suggested that the higher elevation surfaces are older and age decreases with decreasing elevation. The sampling sites at the highest elevation (255-275 m), located on an elongated mesa, yield ages in the 16-12 Ma range. Samples from an intermediate site (225-230 m) yield ages in the 6-4 Ma range. Samples collected at the lowest sites (200-220 m) yield ages in the 2.2-0.8 Ma interval. The age *versus* elevation relationship obtained suggest that the stepped landscapes in the Dugald River area record a progressive downward migration of a relatively flat weathering front. The steps in the landscape result from the differential erosion of previously weathered rock

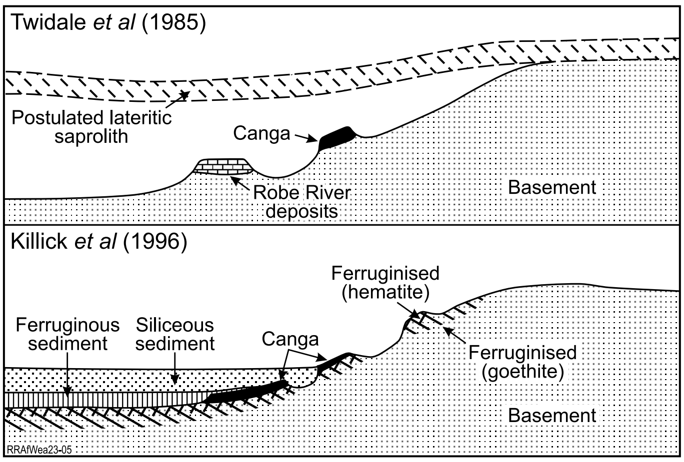

Figure 23. Model of regolith-landscape development for the Hamersley Province proposed by Twidale et al. (1985), compared to Killick et al (1996) (after Killick et al., 1996).

DURICRUSTS AND LAND SURFACES

One of the most important features of the Australian landscape is the widespread occurrence of ferruginous and siliceous duricrusts. These have been used as climatic indicators and have been related to land surfaces, but the models have, however, not been tested. A general view derived from original work by Woolnough (1918, 1927) is that 'laterite' is residual, its formation is restricted to surfaces of a very perfect planation and that there has only been one period of laterite formation. It has been now clearly shown that ferruginization and silicification have affected a variety of residual and transported materials of various ages over long periods and it is not necessarily associated with deep weathering. Ferruginization and silification occurred more than once and, in places, in sites having a marked relief. In places, ferruginous duricrusts form hills because of relief inversion.

No reliable indication of age can be obtained from the morphology, chemical composition and mineralogy of ferruginous duricrusts (Bourman *et al*., 1987). Ferruginous duricrusts with similar morphologies but different chemical and mineralogical compositions can occur at several levels (Figure 24). This is typified by the Buckley River - Lady Loretta area, in the Mt Isa region. The massive and nodular hematitic duricrusts have formed *in situ* on Fe-rich weathered rocks by accumulation of ferruginous material from the mottled saprolite by down-wasting as clays and soluble elements were removed. Slabby duricrust is goethite-rich and is formed on lower slopes by ferruginization of gritty colluvium and upper saprolites; the topography has been inverted since deposition. Vesicular duricrust is goethite- and hollandite-rich and is probably the youngest in the landscape. The Fe oxide mineralogy of the slabby, vesicular, massive and nodular duricrusts suggests that they have developed under different hydrological environments. Free drainage and a high Fe precipitation temperature in the massive and nodular duricrusts yielded hematite; low precipitation temperature, moist conditions and abundant organic matter produced goethite in the slabby and vesicular duricrust. Iron in slabby and vesicular duricrusts is thought to be enriched by absolute accumulation, derived from residual weathering of a local, upland area, and by lateral chemical transport and precipitation in a low landscape position.

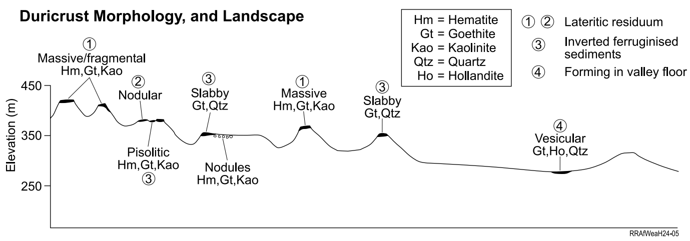

Figure 24. Relationship between ferruginous duricrust and landscape, Mt Isa region (after Anand et al. 1997).

LANDSCAPE AND REGOLITH- EVOLUTION

Landscape evolution models

Cyclic models

The low relief of much of the continent has promoted interpretations of large parts of the Australian landsurface as palaeoplains, within a context of various genetic models, including peneplains (Davis, 1899) and pediplains (King, 1953). The main difference between the two is the concept of *slope decline* of Davis and that of *parallel slope retreat* of King (Figure 25). Many early geomorphological studies favoured the Davisian idea that a vast area of the Australian landscape has evolved through cycles of landscape lowering and relief reduction, punctuated by episodes of landscape rejuvenation. A notable example is the concept of the ‘Great Peneplain’, a single, regionally extensive landsurface across much of the continent (*e.g.*, Andrews, 1903; Woolnough, 1927). On the Yilgarn Craton, interpretations of regional peneplains were suggested by Jutson (1934). Jutson referred to gently undulating uplands as portions of an ‘Old Plateau’ that are capped by deeply weathered profiles, sand and ‘laterite’. He perceived a ‘New Plateau’ developing as a consequence of the erosion of the ‘Old Plateau (Figure 26). In this case, the New Plateau would be free of ‘laterite’. In reconstructing former lateritized landscapes, Jutson and others (*e.g.,* Woolnough, 1918; Stephens, 1946; Prescott and Pendleton, 1952) assumed former continuity of the now isolated, present-day ferruginous duricrust-capped mesas.

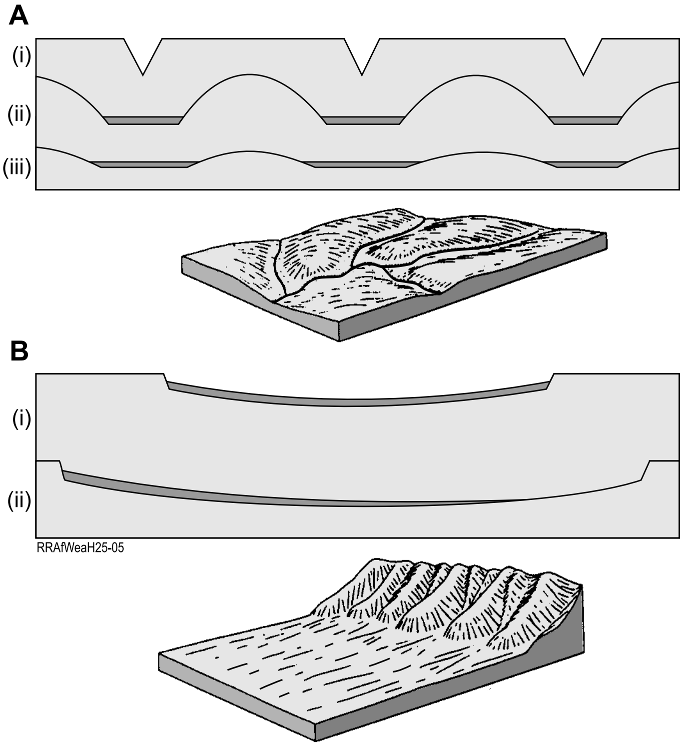

Figure 25. Peneplain and pediplain models of landscape evolution (modified after Thornbury, 1954; Hills, 1975).

These two competing models, scarp retreat and surface denudation, can be tested by study of weathering geochronology (Vasconcelos, 1999). In the Davisian model, surface denudation will progressively remove the products of weathering. If the landscape in the areas studied by weathering geochronology (as discussed above) had evolved in a Davisian fashion, none of the ancient weathering profiles that once blanketed the region would be preserved. However, if a landscape followed a scarp retreat model, remnants of the ancient weathering profile should remain. The geochronological results obtained by the analyses of the Mn oxides and hematite suggests evolution by scarp retreat. This interpretation can be further tested through the application of weathering geochronology to the determination of the time at which the dissected plains became exposed. If the scarp retreat model is correct, weathered material on dissected landscapes should be younger than that on the adjacent plateau, where the ancient regolith is preserved (Vasconcelos, 1999). There has been very little geochronological work in this field, but Mn oxides dates in the Mt Isa and Amazon regions (Vasconcelos, 1995) confirm this hypothesis.

Although extensive sheets of duricrust and gravels do occur, they were never continuous across the landscape. These materials are found at different levels in the landscape and have been formed in a variety of residual and transported materials of diverse ages. Thus, it is questionable to assign a single age to all ferruginous duricrusts (or lateritic residuum) or to imply a single extensive duricrust-capped surface of planation of continental extent. Thus, Jutson’s model of landscape evolution cannot be applied on a regional scale. However, breakaways and mesas capped with duricrusts of similar origin may be correlated on local to district-scales.

Etchplain concept

The etchplain concept (Wayland, 1933) emphasizes the formation of a deeply weathered mantle and its subsequent differential stripping. It was used by Thomas (1965) to classify weathered terrain in Africa and was applied to the SW regions of Western Australia by Finkl and Churchward (1973). At the heart of this concept is the presumption that a continuous weathering profile covered the entire area now classed as the etchplain. Differential degrees of stripping of the deeply weathered mantle are recognized by the use of such terms as incipient, partial, semi-stripped and stripped plains (Figure 27) (Finkl, 1979). Stripping of regolith materials, as shown in Figure 27, progresses towards the interior in a stepped fashion from the N and S.

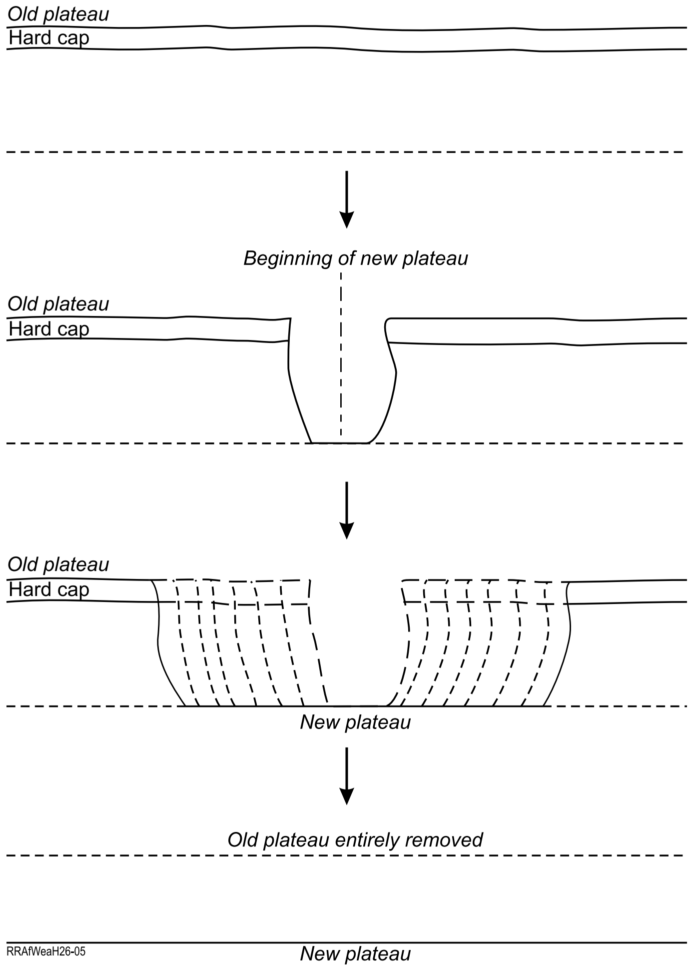

Figure 26. Jutson’s diagrammatic explanation of the formation of the New Plateau from the Old Plateau (after Jutson, 1934).

Relief inversion model

More recently, Ollier *et al*., (1988) and Ollier and Pain (1996) suggested a very different interpretation for the evolution of the weathered landscapes. They concluded that weathered material has been repeatedly deposited in valleys, after which the relief became inverted, so that relicts of alluvium occur as a ferruginous duricrust capping (Figure 28). If such inversion of relief has occurred, a totally different landscape history emerges, and certainly the regolith history will be much more complex (Ollier and Pain, 1996). However, although there are numerous sites showing local-scale relief inversion in deeply weathered terrains of Australia, there is little evidence for regional-scale inversion and the concept has been questioned by several authors (*e.g.*, Conacher, 1991, Davy and Gozzard, 1995). It is relevant to note that in the Yilgarn and Gawler Cratons, for example, major drainages, some filled with Permian sediments, still occupy lower parts of the present landscape, and there are no thick sedimentary units in elevated areas. Where local relief inversion has occurred in areas of duricrust, it is all too easy to cast phantom ‘peneplains’ across the summit remnants, and come up with a story of successive erosion surfaces.

Regolith evolution

Despite glaciation during the late Proterozoic and Permian, some traces of earlier regolith and landforms remain. Deep kaolinitic weathering

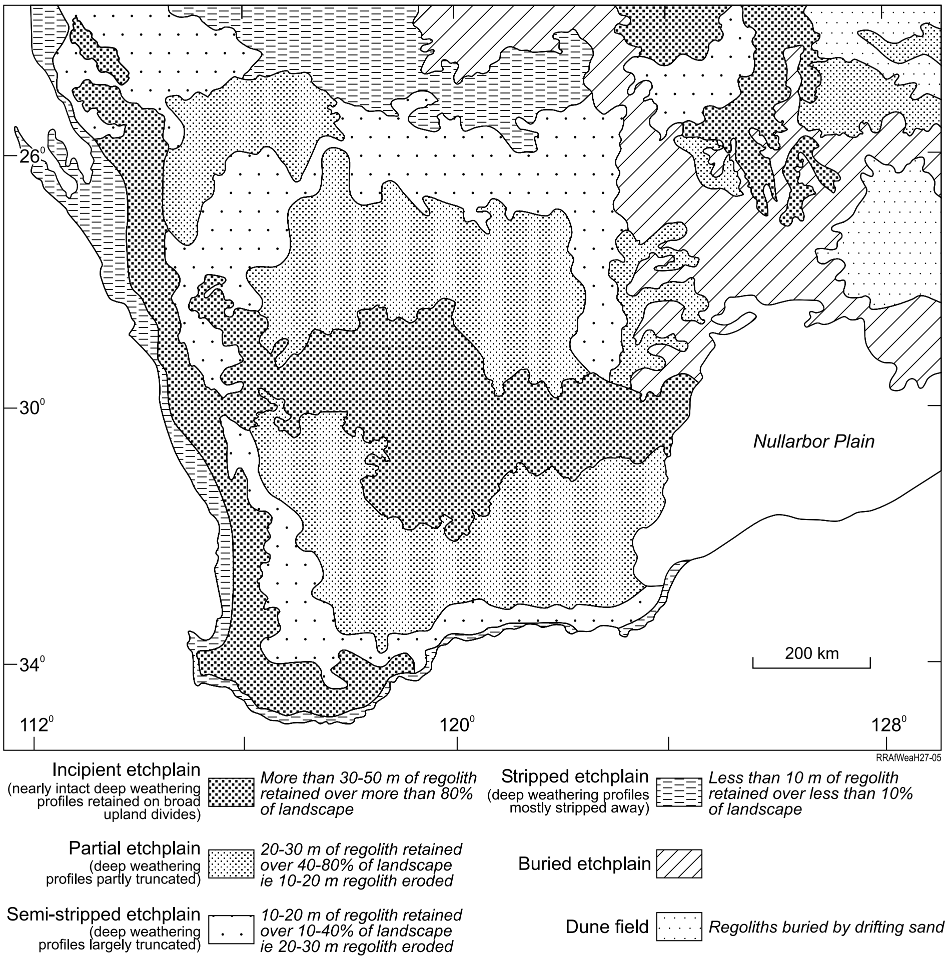

Figure 27. Distribution of etched surfaces in southwestern Australia (after Finkl, 1979).

commenced in the Late Cretaceous – Palaeocene or even earlier and continued to mid- to late- Miocene. Dissection of the profiles has generally been shallow and has not proceeded below the saprolite, particularly in much of the Western Plateau and basins. Ferruginization, silicification and calcification have affected a variety of residual and transported materials over long periods; ferruginization and silicification have clearly occurred more than once (Figures 29 and 30). Calcification is currently active in arid areas, but evidence for earlier episodes is equivocal. Biological processes appear to have played important role in the formation of ferruginous materials. Evidence of ferruginization from the Pre-Eocene through to at least Pliocene can be seen in the arid parts of the Yilgarn Craton and Girilambone-Cobar region. Alley *et al* (1999) reported that ferricrete in south-central Australia probably formed in early Mesozoic, Late Oligocene-Middle Miocene and Late Miocene-Pleistocene. In the Yilgarn Craton, Eocene sediments contain detrital pisoliths apparently eroded from ferruginous duricrust, indicating that it formed before the end of the Eocene (Late Cretaceous

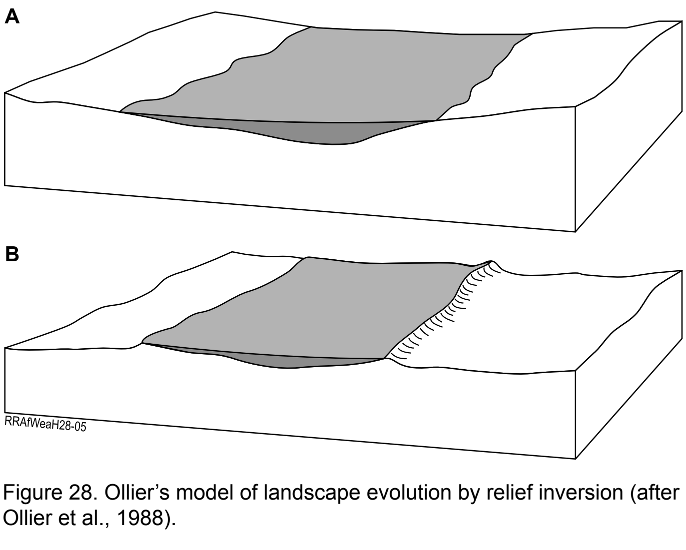

Figure 28. Ollier's model of landscape evolution by relief inversion (after Ollier et al., 1988).

and Palaeocene). Evidence for later ferruginization lies in weathering profiles developed in sediments, including those in inset-valleys. Some of these sediments were deposited in the late Eocene and have themselves have been strongly weathered, with strongly mottled and pisolitic upper horizons and cemented duricrusts. Palaeomagnetic data suggest that the second major phase of ferruginization appear to have occurred in Middle to Late Miocene and the youngest in Pliocene. The transport of the Fe to form a variety of ferruginous duricrusts in inland areas is related to past humid episodes rather than recent arid climates in which Fe transport is insignificant. However, ferruginization is still active in south western WA (*e.g.,* the Darling Range), northern NT, N E Qld and other humid areas.

The duration of episodes of ferruginous duricrust formation is unknown. Mann and Ollier (1985) demonstrated that a 1 m thick Fe oxide-rich duricrust could form in 10 000 years by ionic diffusion, but they did not suggest that this actually occurred. Field evidence indicated much greater ages for some, probably a million years (Ollier and Pain, 1996). It has also been suggested by Nahon (1986) that a ferruginous duricrust, a few metres thick, which overlies tens of metres of saprolite, needs several million years to form. Recently, Theveniaut and Freyssinet (1999) used palaeomagnetic data to interpret the succession of horizons, from the weathering front to the top of the duricrust surface, of a thick bauxitic profile in French Guiana. The duricrust is divided into two levels within the massive facies; the base of duricrust shows palaeomagnetic directions that fit with the present-day situation, whereas the upper part records directions of the South American pole wander path older than 10 Ma. These authors concluded that hematite precipitation was not uniform within the duricrust during 'lateritization', and seems to preserve traces of several successive and distinct phases.

There are several widespread episodes of silicification. Although silcrete is highly variable, there are two main types: groundwater silcrete, which retains primary rock fabric and structures, and pedogenic silcrete, which displays vertical differences in structure and silica mineralogy (Wopfner, 1978; Thiry and Milnes, 1991). There have been at least two episodes of silcrete formation and one of red-brown hardpan

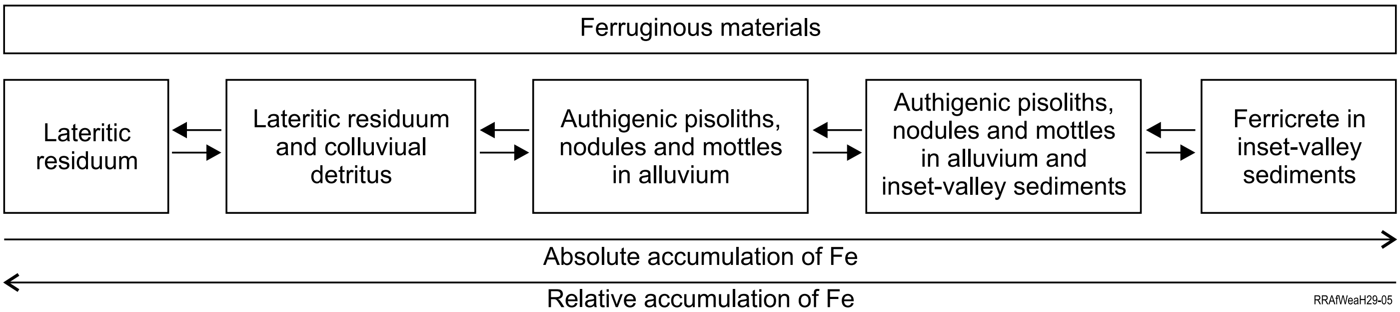

Figure 29. Classification of ferruginous materials illustrating the continuum between lateritic residuum and ferricretes.

formation, but the timing is unclear. Wopfner (1978) and Idnurm and Senior (1978) suggested that the main period of silicification was generally after deep kaolinization, *i.e.,* Eocene-Miocene. According to Alley *et al.* (1999), major phases of silicification occurred in Late Eocene-Middle Eocene and Late Miocene-Pleistocene when significant groundwater silcretes formed. Because, in places, silcrete pre-dates ferruginous duricrust, it is not possible to invoke Late Tertiary aridity to account for its formation. The silcretes in the Tick Hill area in the Mt Isa region are at least 38 Ma old, indicating that these silcretes are not associated with postulated transition towards arid conditions in the Miocene (Vasconcelos, 1998). As silicification requires the contemporary release of silica by chemical weathering, an adequate rainfall is indicated. Thus, a humid tropical or sub-tropical environment of low relief was suggested by Summerfield (1983). The acid conditions necessary to dissolve kaolinite would be promoted by degraded organic matter and ferrolysis reactions in poorly-buffered siliceous soils. However, precipitation, whether at the surface or at depth, requires concentration, presumably by excess evaporation. Accordingly, it seems that silicification probably occurred during transitional periods between humid and arid phases between the Late Eocene to Oligocene, with silica accumulating in the groundwaters as drainages became less competent, with precipitation at seepages, water-tables and porosity barriers. Such a scenario was suggested by Butt (1985) for the formation of pedogenic silcretes formed on deeply weathered granitic rocks in the northern Yilgarn Craton. Hill (2000; 2005) suggested that silcrete development in some Mesozoic and Cainozoic sedimentary sequences could be due to acidic weathering conditions associated with organic acids and pyrite weathering. This would lead to loss of alumina and local redistribution of silica. A second major phase of silcrete formation occurred in the Pleistocene. Ferruginization of silcrete is widespread and ranges from a brown colouration (mottling) to Fe-cementation of siliceous nodules. If silcretes represent evidence for supergene silica precipitating during arid conditions, ferruginization and replacement of silica suggest a later humid phase.

Quaternary climate changes have been important in the development of regolith and landforms, particularly affecting colluvial, alluvial, aeolian and lacustrine sediments. With the change to aridity in the Late Miocene or Pliocene (Figure 30), soil processes became dominated by calcification and silicification rather than ferruginization. Chemical inputs associated with atmospheric contributions (rain and dust) are important control on the widespread development of pedogenic calcrete in winter rainfall-dominated areas in the S that are more proximal to marine Ca and Mg sources than the summer-dominated rainfall areas that are more distal to marine Ca and Mg sources.

Extensive hardpanization (silicification) of younger sediments and residual regolith indicates widespread mobilization of silica. The cementation still appears to be active particularly in the central arid zone. Calcification is the youngest event and postdates the onset of silicification. Precipitating carbonates, amorphous silica and aluminosilicates may displace and replace the host matrix, including saprolite and ferruginous duricrust, whether *in situ* or transported.

With increased aridity, surface drainage channels have become choked with sediment and groundwaters in many areas, especially in the S of the continent, have become saline. The major valleys became the sites of chains of salt lakes, with aeolian processes forming adjacent gypsum dunes, lunettes and lake parnas. Aeolian deflation also left vast areas covered by polished, ferruginous, siliceous, lithic or polymictic lag, which, in some places, forms gibber plains; the fine materials contributed to soils elsewhere. Lithosols are associated with fresh rock or saprock and areas of steeper slopes. Soils contain abundant calcrete and smectite from recent weathering of mafic rocks.

Recent overprinting of ferruginization has continued in arid areas. Inland, biogenic-related processes, calcification or silification appear to have triggered deferruginsation of duricrusts and mottled zones. The processes of calcification and silicification have clearly resulted in removal of Fe, possibly by replacement, but the mechanisms are not well understood. In humid regions, degradation of bauxitic duricrust continues by dissolution of goethite and gibbsite at or near the surface. Similarly the present climate provide a favourable environment for the strong leaching and active formation of gibbsite under the winter rainfall of the Mediterranean climate.

IMPLICATIONS FOR EXPLORATION

Economic significance

The complex regolith environments both challenges and provides opportunities for mineral exploration. Firstly, the regolith presents numerous exploration problems, affecting geological, geophysical and geochemical mapping and exploration techniques, and constraining their use. The presence of transported overburden of various ages exacerbates these problems, especially for geochemical procedures based on surface or near-surface sampling. Geophysical survey methods, particularly airborne techniques, are commonly applied to see through this cover, but are hindered by interfering regolith-related responses. Conversely, dispersion during regolith formation and evolution potentially yields much larger exploration targets than the mineralization itself. In addition, the regolith itself may represent economically significant mineral deposits:

- Weathering products of otherwise 'unmineralized' rocks: bauxites, some Fe and Mn ores, Ni-Co laterites, and various industrial minerals, including clays, such as kaolinite. These include some of Australia's major mineral resources, including bauxites at Darling Range, Weipa, Gove and Mitchell Plateau, and Mn at Groote Eylandt. Demonstrated resources of bauxite in Australia are estimated at more than 8300 Mt, of which at least 3000 Mt are economic demonstrated resources, the highest in the world (Lambert and Perkin, 1998). The Darling Range is the world's leading alumina producing region. In Western Australia, Ni laterite resources amount to some 600 Mt of Ni metal and represent a major laterite resource in the world. The Ni is hosted primarily by silicate minerals (silicate-smectite type deposits) including Bulong and Murrin Murrin, or by Fe (and Mn) oxides (oxide type deposits) (Brand *et al.,* 1996; 1998).
- Weathering resistates and placer deposits; mineral sands, diamonds, alluvial gold. Terrestrial deposits are mostly minor, although mineral sands in old marine terraces and shorelines are, ultimately, weathering resistates. Rare earth elements and apatite sands, accumulated over carbonatites, (*e.g.,* Mt Weld Range, Yilgarn Craton) may be a mixture of primary and secondary minerals.
- Secondary enrichment of primary mineralized systems: supergene Au, Cu and U deposits. Most important of these are the numerous Au deposits found in Western Australia, the northern Territory and, increasingly in eastern Australia (*e.g.*, Mt Gibson, Boddington,

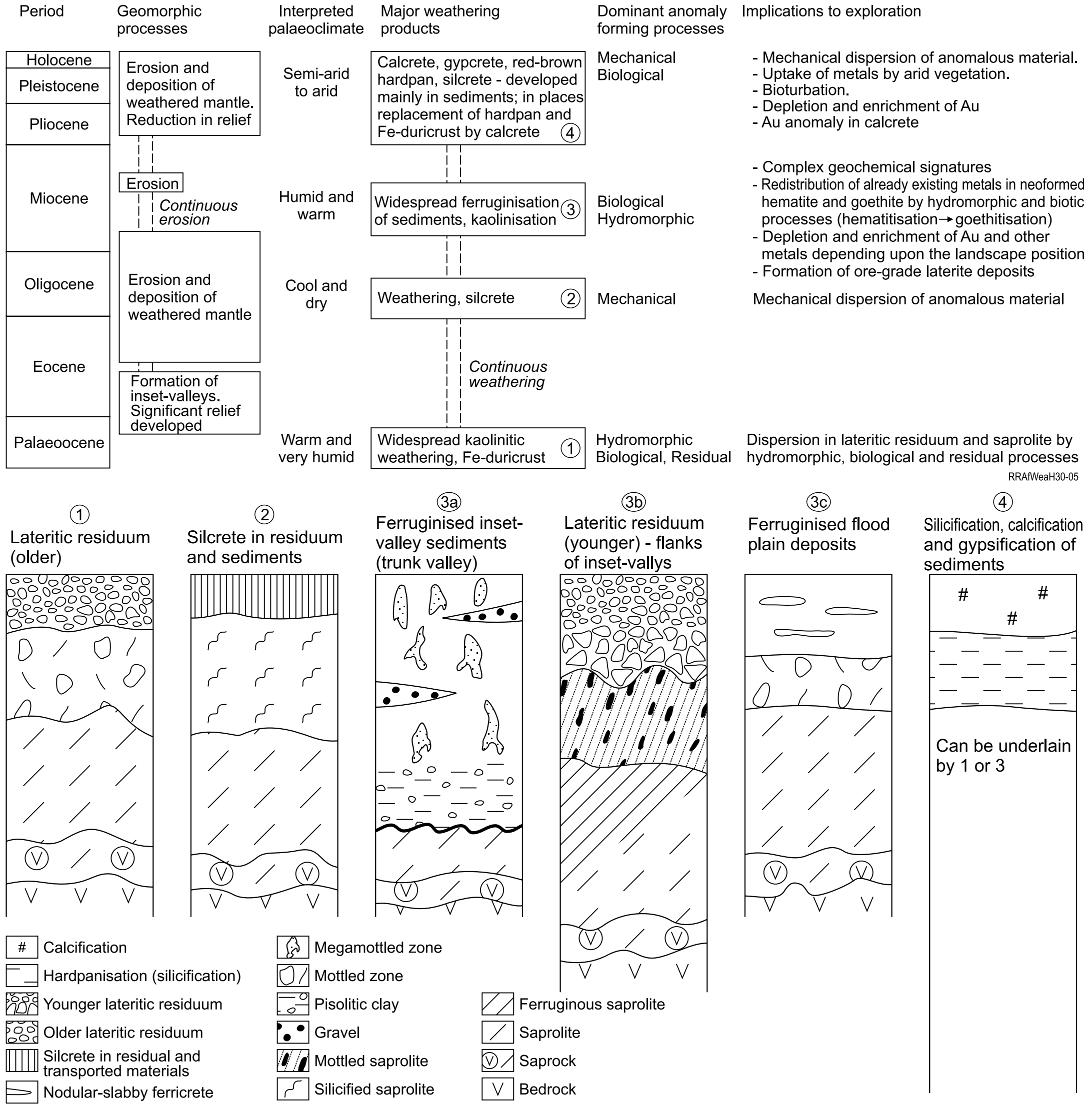

Figure 30. Generalized weathering history of the Yilgarn Craton and its implication for exploration. Weathering and erosion have been continuous, but their rates and relative importance changed over time in response to climate and tectonic events.

Paddington, Tanami, Northparkes). Some of these developed on significant primary mineralization and others, mostly smaller operations, where the primary source is uneconomic (Butt, 1998).

- Epigenetic enrichment in sediments: Fe, U and, possibly, some Au deposits in inset-valley sediments and shallow basins. Iron ores, such as the Robe River deposits, and both oxidized (*e.g.*, Yeelirrie) and reduced (*e.g.,* Lake Frome Basin) styles of U deposit have formed by precipitation in redox-controlled trap sites in sediments, distant from a dispersed source.

Two and 3D regolith-landform mapping

The formation of regolith, its modification under changed climatic settings and the effects of erosion during and after these events results in significant regolith-landform control on sample media. Knowledge of the distribution and properties of regolith materials is essential for successful exploration in regolith-dominated terrains, whether for deposits concealed by the regolith or for those hosted by it. Regolith-landform mapping is an essential first step, followed by characterization of the regolith materials themselves. Ideally, regolith and landform maps should be produced at the start of an exploration programme. These can include factual maps, which show the units present with little or no genetic bias, and interpretative maps and models that summarize the most appropriate geochemical sampling strategies for each terrain unit.

An essential feature of regolith-landform mapping is the concurrent development of a model(s) of regolith and landform evolution, and relating these to exploration models (*see* Models, *this volume*). These models are developed from analysis of a broad regolith-landform framework of an area including elevation distribution, location of major landform features, drainage pattern analysis and regolith studies at a detailed scale. Such models reflect an understanding of the distribution of regolith materials in relation to landforms and weathering history. Their predictive qualities can lead to a more efficient and effective mapping and exploration programs. Satellite imagery and aerial photography remain the basic approaches to mapping, but are routinely supplemented by multispectral Landsat and, increasingly, airborne hyperspectral data for mineralogical information, and airborne radiometric surveys for geochemical data. However, whilst these procedures indicate the nature and distribution of surface materials, they are poor indicators of the underlying regolith, especially in areas of transported overburden and basin cover. Existing 3D models to date have been built around large drill-hole datasets. These models, which define the base of the transported regolith, weathering front, and the internal architecture and distribution of inset-valleys, have genuine predictive value. However, they are difficult to construct even from good datasets due to inconsistencies in logging, and the difficulty in actually defining the unconformity between transported-residual materials. Low-frequency, broad-bandwidth, time-domain airborne EM systems show promise map for mapping regolith thickness and architecture, with a minimal amount of drill-hole calibration.

For devising exploration strategies, it is important to develop prototype derivative maps showing residual areas, transported material less that 5 m thick, and areas deeper than 5 m. These maps are created by fitting a form surface based on the extrapolation of the residual areas under major areas of transported regolith, guided by digital terrain and landform models, together with high-resolution aeromagnetic surveys and drilling (McQueen, 2004). These techniques require considerable development before they can be considered accurate and rapidly produced tools for the mineral explorer.

Preliminary fieldwork in all regolith-landform regimes should include deep pitting or drilling to provides a basis for selecting sample media and interpreting data; it is essential where there is transported cover. Even simple observations, such as the colour of termite mounds, can indicate whether soil is derived from saprolite or transported cover, crucial to determining sampling strategies. Logging, however, may not be easy, even where there is good field exposure, and can be very difficult from drill cuttings, yet correct identification of regolith materials and boundaries (*e.g.*, transported vs residual) is of considerable importance. For example, ferruginous materials are particularly valuable as sample media but, as noted above, there are numerous types of these materials (Figures 29-30) whose relationship to bedrock and mineralization vary quite considerably. It is expected that the development of instrumental procedures for routine logging will be of great value in overcoming many of these problems.

ACKNOWLEDGEMENTS

The author would like to thank Drs Charles Butt, Ian Robertson and Matthias Cornelius for reviewing the manuscript.

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