DISTRIBUTION AND EVOLUTION OF 'LATERITES' AND LATERITIC WEATHERING PROFILES, DARLING RANGE, WESTERN AUSTRALIA

R. R. Anand and C.R. M. Butt
CRC for Landscape Environments and Mineral Exploration, CSIRO Exploration and Mining, 26 Dick Perry Avenue, Technology Park, Kensington, Western Australia 6151.

ABSTRACT

This paper summarizes the distribution and characteristics of the lateritic regolith of the Darling Range and presents models for its formation and evolution. Typical, complete, weathering profiles on granite average about 20 m in thickness and consist of gravelly soil, lateritic duricrust, saprolite and saprock. Lateritic duricrusts occupy gently sloping to horizontal upland areas and are either residual, or locally transported and recemented. In much of the Darling Range, the lower part of the duricrust, especially on hill slopes rather than crests or valley floors, is highly aluminous and forms an extensive resource of bauxite. Fragmental, fragmental-pisolitic, pisolitic and vesicular types can be identified on the basis of secondary structures. Fragmental duricrust largely consists of gibbsite, hematite, goethite and quartz and has resulted from direct gibbsitization of saprock or bedrock without forming the kaolinite-rich deep saprolite. Outcrops of duricrust with relict bedrock textures are common. In contrast, pisolitic duricrust with hematite-maghemite and χ-alumina rich mineralogy have a more complex history than fragmental duricrust with simple mineralogy. Vesicular duricrust is goethite-rich and is formed by the ferruginization of sandy detritus and quartz pebbles. The profiles show no condensed sequences, and individual rock types are traceable geochemically and mineralogically, but with increasing difficulty, from bedrock to the surface. The concentrations of Fe, Al, Si, Ti, V, Cr and residual quartz, particularly in fragmental duricrust, can be used to identify bedrock. Deep weathering profiles at Jarrahdale and Boddington yield late Tertiary palaeomagnetic ages and it appears that modification of these profiles to form bauxite is continuing today.

1 INTRODUCTION

The Darling Range is a dissected, uplifted plateau developed on the western margin of the Archaean Yilgarn Craton and some adjacent Mesozoic rocks of the Perth Basin. It extends from approximately the Moore river 100 km north of Perth to Collie and Bridgetown, 220 m south. To the west, it is bounded by an escarpment, which approximately follows the Darling Fault, separating the Yilgarn Craton and the coastal plain. The eastern boundary is less clearly defined, but in part follows the lowlands, including the Avon valley, marking the Cape Riche-Yandanooka lineament. A deeply weathered lateritic regolith is developed throughout the range, on igneous and metamorphic Archaean rocks and sediments of Permian and Mesozoic age. The regolith, which varies in thickness from a few metres to over 50 m, is generally well preserved, except in the valleys of the major drainages that pass through the range (Swan-Avon, Murray, Collie and Blackwood rivers). It is also being eroded by smaller streams and rivers that rise in the Range itself. From east of Perth to the Harris River area, north of Collie, the regolith is bauxitic and forms the world's leading alumina producing region.

Lateritic duricrusts are characteristic and important features of the Darling Range. They occupy gently sloping to horizontal uplands and are considered to have formed by *in situ* weathering of parent rocks (Baker, 1972; Sadleir and Gilkes, 1976; Davy, 1979; Murray, 1979; Butt, 1981; Davy and El-Ansary, 1986; Anand et al., 1991; Hickman et al., 1992; Anand, 1994; Anand and Paine, 2002) and weathering of fluvial (Grubb, 1971) or aeolian sediments (Brimhall et al., 1988; Glassford and Semenuik, 1995). There have many discussions on the role of parent rock, climate, topography, tectonic history and vegetation in the development of bauxitic profiles (Terrill, 1956; Playford, 1959; Grubb, 1971; Baker, 1972, 1976; Sadleir and Gilkes, 1976; Davy, 1979; Hickman et al., 1992). The objectives of this paper are to: (i) summarize the distribution and characteristics of regolith materials in the Darling Range including lateritic duricrusts and (ii) present models of their development and evolution.

2 CLIMATE AND VEGETATION

The Darling Range has a typical, highly seasonal, Mediterranean climate, with cool, wet winters and warm dry summers. The mean maximum temperature ranges between 12 and 28°C. The total rainfall is closely related to latitude, distance from the coast and local relief and elevation. Rainfall ranges from over 1300 mm per annum on high ground (>300 m AHD) near the scarp to below 460 mm per annum in the Avon valley near York. The grade of bauxite decreases progressively with the present day rainfall of the area so that, where rainfall is less than 600 mm, bauxite grade is below the level of commercial exploitation. The dominant vegetation is open eucalypt forest. In wetter areas, near the scarp, varying proportions of *Eucalyptus marginata* (jarrah) and *E. calophylla* (marri) form the upper storey on the ridge crest and slopes. In drier areas inland, the woodlands are more open and characterized by abundant *E. wandoo* and *E. accedens* (wandoo). *Banksia grandis, Persoonia longifolia*, *Xanthorrhoea preisii* and *Macrozomia riedlei* dominate the middle and lower storey. *E. calophylla* and *Melaleuca preissianan* dominate the upper storey of the valley floors, where the lower storey contains *Xanthorrhoea preisii*, *Kingia australis* and *Hypocalymma angustifolium.*

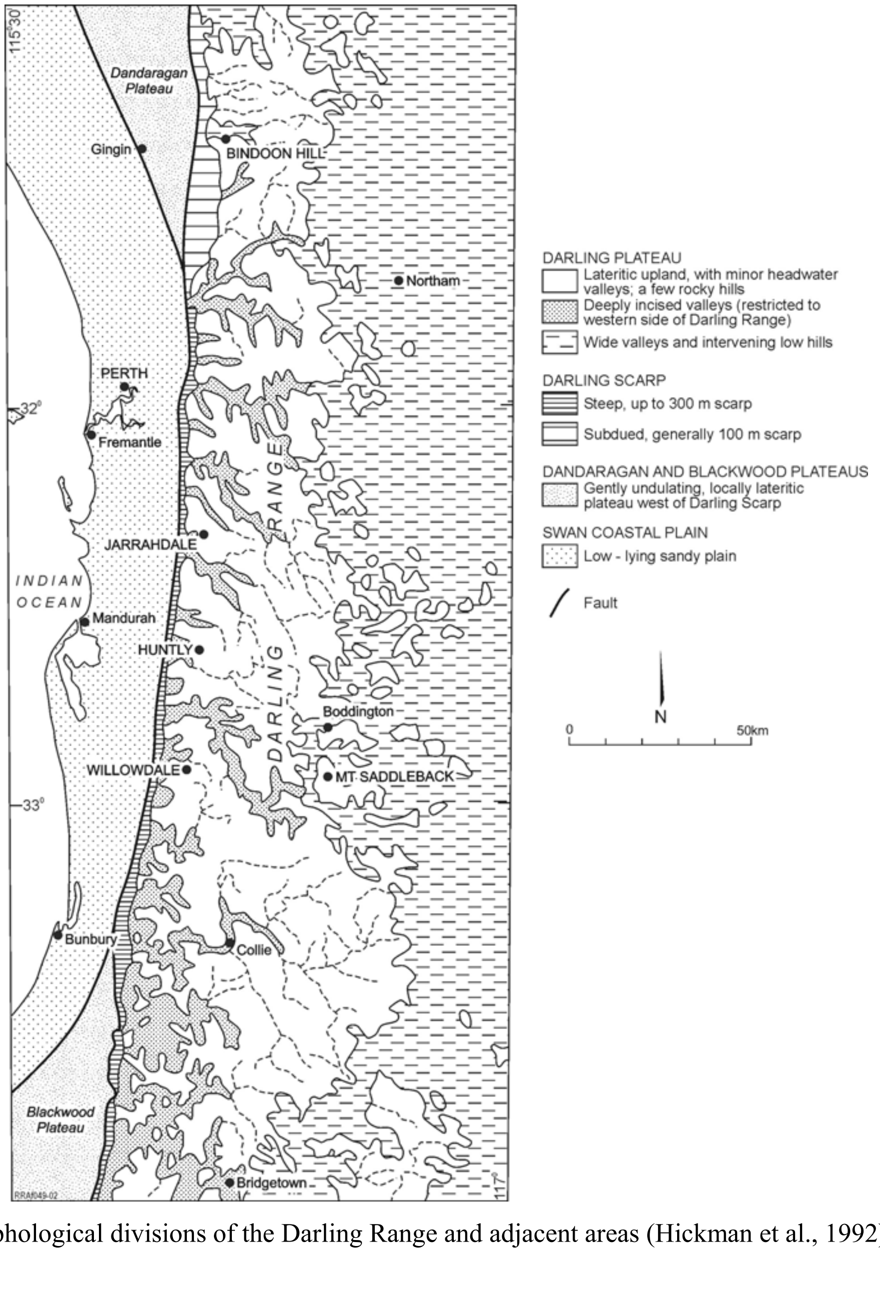

Figure 1. Geomorphological divisions of the Darling Range and adjacent areas (Hickman et al., 1992).

3 BEDROCK GEOLOGY

Much of the Darling Range is underlain by crystalline rocks of Western Gneiss Terrane of the Archaean Yilgarn Craton. The rocks are principally granites and granitic gneisses that have been intruded by numerous northwest trending late Proterozoic tholeiitic, quartz dolerite dykes that commonly constitute over 10% of bedrock (Hickman et al., 1992). The oldest units (>3.0 Ga) are the Jimperding Gneiss Complex and Chittering Metamorphic Belt in the north, and the Balingup Gneiss Complex in the south. These consist dominantly of orthogneiss and paragneiss, with metamorphic assemblages of amphibolite to granulite grade. The remaining granitic terrain is younger, 2.55-2.7 Ga, and of low metamorphic grade, and has intruded mostly older (2.65 Ga) greenstone rocks of the Saddleback Group. The Saddleback Group extends for 43 km north-northwest from near Boddington and is between 5 km and 12 km wide. It consists of a weakly metamorphosed felsic and mafic volcanic and pyroclastic rocks, with minor sedimentary units, largely fault-bounded against granite and migmatite.

Post-Proterozoic sedimentary rocks in the Darling Range include coal-bearing Permian sandstones and siltstones that occupy small basins in the Archaean rocks at Collie and Wilga, and up-faulted Mesozoic sediments of the Perth Basin that form the Dandaragan Plateau, north of Perth, between the Gingin scarp, which follows a western splay of the Darling Fault, and the Darling Fault itself. There are also small but widespread occurrences of deep unconsolidated quartzose sediments, sometimes containing rounded siliceous gravels (up to 7 m thick). These may occur along shallow valleys or valley divides as well as on low crests (Mulcahy et al., 1972).

4 GEOMORPHOLOGY

The Darling Range has been divided into three geomorphological provinces (Churchward and McArthur, 1980; Hickman *et al.,* 1992), namely the Dandaragan Plateau in the north. the Darling Plateau and the Blackwood Plateau in the south (Figure 1). This discussion is confined to the Darling Plateau.

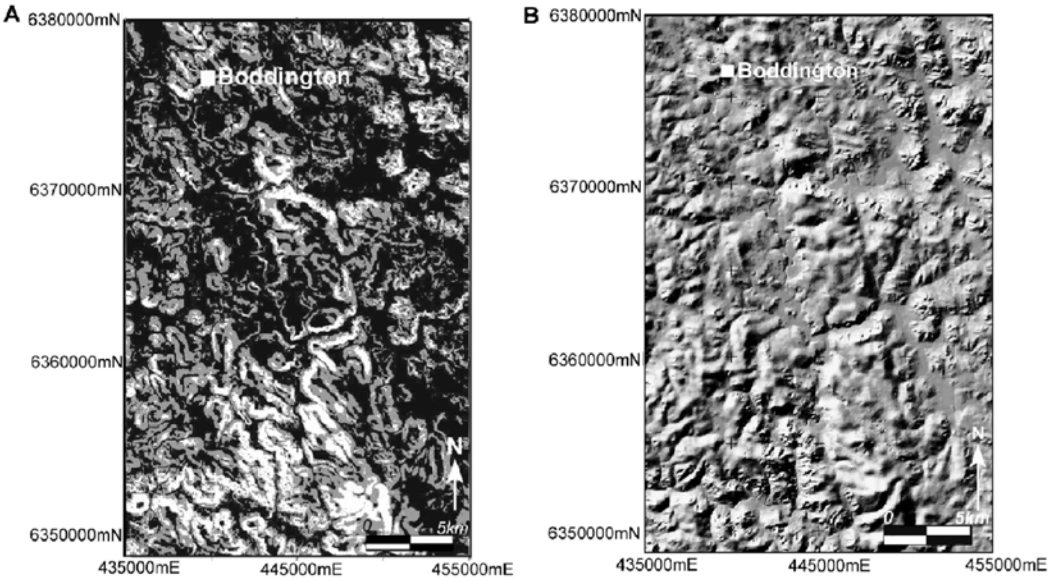

Figure 2: Geomorphological features of the Darling Range. (A) Intensity of slope about the Boddington district with slopes ranging from 0° to 35°. (B) Shaded elevation model about the Boddington district (data courtesy of the Australian Surveying and Land Information Group, Canberra, Australia. Crown Copyright©. All rights reserved).

The western boundary of the Darling Plateau, the Darling Scarp, rises abruptly 100 to 300 m above the low-lying subdued topography of the Swan Coastal Plain. The position of the scarp closely corresponds to that of the Darling Fault, which separates resistant Precambrian rocks to the east from Phanerozoic sedimentary rocks of the Perth Basin. The Darling Plateau has an undulating surface (e.g. Boddington district, Figure 2A, B), with crests ranging in elevation from 280 to 330 m and swampy valley floors set some 50 to 100 m below. Projecting above the general plateau level are summits rising to over 500 m, the most prominent of which are Mt. Cooke (582 m), Mt. Dale (543 m), Mt. Randall (523 m), Mt. Solus (576 m), Wourahming Hill (503 m) and Mt. Saddleback (575 m). The terrain is dissected in the east and more so in the west, a trend that relates to progressive drainage incision from shallow swampy tracts on the plateau to deep valleys near the Darling Scarp. Dissection of the plateau has created three distinct subdivisions: lateritic uplands, deeply incised valleys and wide valleys with intervening low hills (Figure 1).

The *lateritic uplands* are mantled with deeply weathered profiles capped with lateritic duricrust, lateritic gravel and sand and form a undulating surface; sand generally occurs in shallow depressions at the heads of drainages. *Deeply incised valleys* are restricted to the western part of the plateau where streams and rivers have cut narrow, steep-sided valleys through the surface of the lateritic upland. Erosion is commonly sufficiently deep to expose bedrock in the lower slopes and valley floors. *Wide valleys* and intervening areas of low hills occur in the eastern part of the Darling Plateau. Small areas of poorly-cemented lateritic duricrust occur on interfluves; bedrock is locally exposed in the more hilly areas (Hickman *et al*., 1992). Topography is generally subdued although deep valleys are associated with major drainages such as the Avon, Murray and Collie rivers and their larger tributaries.

5 WEATHERING PROFILES

The interplay of weathering, erosion and relocation of materials, generally over long periods, has formed the lateritic regolith that mantles the Darling Range. On granite, the regolith (Figure 3) commonly averages about 20 m in thickness and consists of gravelly soil, lateritic duricrust (with a lower, bauxitic, zone in wetter areas), saprolite and saprock. Similar profiles form on greenstones, but are generally deeper (>40 m). However, in detail, profiles are extremely variable – for example, considerable thicknesses of kaolinitic clay form where porosity is greater over shears and faults. In places, transported duricrust overlies fresh bedrock.

Soil forms the upper part of the regolith. It is generally 0.2-2.0 m thick (mean ~0.5 m) and consists of lateritic gravel in a clayey sand matrix. The soil overlies lateritic duricrust, which reaches 1-3 m in thickness over granite and up to 4 m over mafic rocks. Over granitoids, the duricrust is light-brown, with abundant visible quartz; over dolerite dykes, it is red-brown and contains very little quartz. Pisolitic duricrust forms the upper unit of the duricrust and consists of black to red-brown, simple or compound, nodules and pisoliths cemented by gibbsite and goethite. This unit may be either essentially residual, or locally transported and cemented. The underlying fragmental duricrust is commonly residual, as indicated by clasts that preserve primary lithic fabrics, such as the original quartz grains and feldspar lathes pseudomorphed by gibbsite of granitoid bedrock. The friable bauxite zone immediately underlies the fragmental duricrust. This zone is yellowish to reddish brown, generally about 3 m thick but can be up to 10 m thick in mid slope positions on large ridges. The bauxite zone contains round to angular nodules and saprolitic fragments in a fine grained, loose, earthy or sandy matrix. The coarser material ranges from 2 mm to over 100 mm. Feldpars pseudomorphs are common in gibbsite nodules and are generally 1-2 mm in size.

The contact between the bauxite zone and saprolite is commonly abrupt. The saprolite is generally 10-15 m thick over granitoids, generally mottled by Fe oxides in the upper part, but becoming increasingly bleached with depth. It consists of vermicular kaolinite pseudomorphs after primary grains of mica, an isotropic groundmass of kaolinite after feldspar, coarse quartz and abundant large voids. The saprolite thus has a much lower bulk density and higher porosity than the parent rock. Over mafic rocks, saprolite ranges from white to multicoloured kaolinitic clay with mottles, commonly overprinted by liesegang rings. Saprolite passes downwards into saprock, before fresh bedrock is reached. The weathering front on dolerite commonly occurs over a few centimetres, whereas over granitoids, it is generally gradational over 2 - 3 m. On a local scale, the topography of the weathering front is much more irregular than that of the duricrust. Pinnacles of bedrock and isolated corestones may occur high in the weathered profile.

On steep slopes, the lateritic profile may be partly or completely eroded, or have never fully formed. In such situations, there may be a surficial transported unit, commonly containing lateritic gravels derived from upslope, overlying weathered or fresh bedrock. The soil is developed from the transported material or, in its absence, directly from saprolite or bedrock.

The economically significant bauxite zone is not always present even where the regolith is fully preserved. It is best developed in freely drained slopes, becoming less aluminous (i.e. not bauxitic) on crests, lower slopes and valleys. As noted above, it also has formed only in higher rainfall areas of the Range between Perth and Collie. Farther south, east and north, it is less abundant. However, an exception to the overall pattern of bauxite becoming less common eastwards is the anomalously high bauxite concentration developed on the Mount Saddleback greenstone belt, where bedrock composition and topographic conditions have combined to produce locally favourable conditions for bauxite formation.

6 MINERALOGY OF WEATHERING PROFILES

The mineralogical compositions of bauxitic laterite profiles developed on granite, andesite and metabasalt are compared in Figure 4. Iron oxides and gibbsite are the dominant minerals in the lateritic duricrust and bauxite zone, whereas kaolinite and/or halloysite dominate the saprolite. Quartz is present throughout the regolith on granitoids. Goethite, hematite and maghemite abundances generally increase towards the top of the profile. They may be very heterogeneously distributed, being absent from certain horizons, yet being dominant minerals only a few centimetres below. They are more abundant on mafic than in granitic profiles. Goethite may occur anywhere in the profile, but is especially common in the upper horizons. On granite, the abundance of goethite generally decreases sharply at the lower contact and is absent from the underlying white saprolite. Where the profile is derived from mafic rocks, as at Boddington, goethite persists though the saprolite. Hematite, although much less abundant is nevertheless common, particulary in the duricrust. Maghemite is present in near-surface pisolitic duricrust and loose pisoliths, commonly with traces of corundum, associated with relatively high concentrations of hematite and low concentrations of kaolinite. The maghemite-rich pisoliths contain large amounts of Al_2O_3 (40-50%), which indicates that maghemite probably did not form from solution, since this is inhibited by dissolved Al (Taylor and Schwertmann, 1974; Coventry et al., 1983. Maghemite and corundum can be formed by dehydroxylation of Fe and Al oxyhydroxide by heating to approximately 300-500°C in the presence of organic matter (Anand and Gilkes, 1987a). Field relationships, including the association of hematite, maghemite and corundum, suggest this to be the probable mechanism, with suitable conditions provided by ancient forest fires (Schwertmann and Fechter, 1984; Milnes et al., 1985; Anand and Gilkes, 1987a)

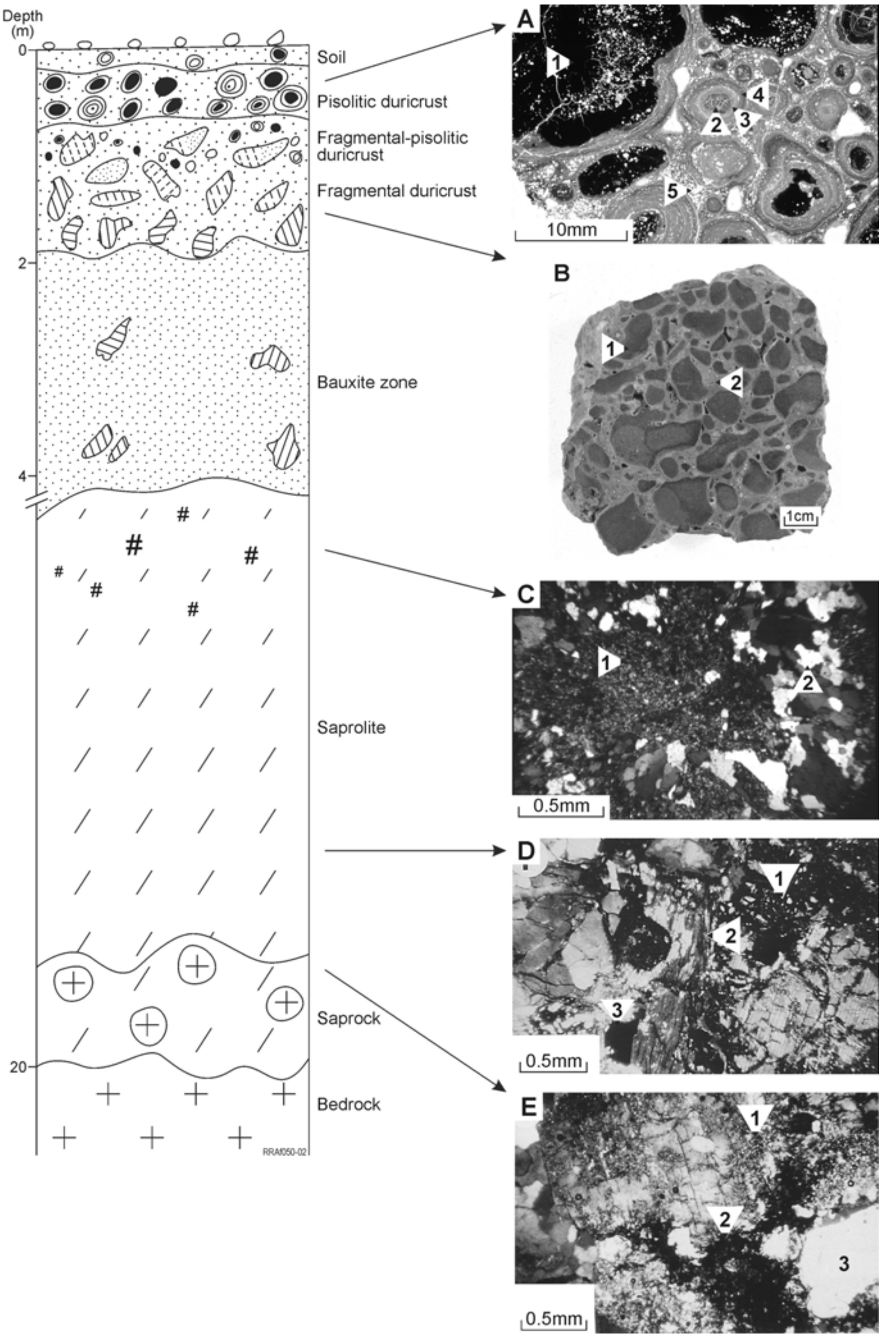

Figure 3: Typical lateritic bauxite weathering profile. Photomicrographs of fabrics from various horizons of a profile developed over granite at Jarrahdale. (A) Optical micrograph of a polished section of pisolitic duricrust showing hematite-maghemite-rich (1) and hematite-rich pisoliths (2) surrounded by lighter (3) and darker (4) cutans in a quartz-gibbsite-rich matrix (5). (B) Fragmental duricrust showing angular to sub angular hematite-gibbsite-rich fragments (1) in a gibbsite-quartz-rich matrix (2). (C) Optical micrograph of upper saprolite showing feldspars completely replaced by kaolinite (1) set in a close packed relict quartz (2). (D) Optical micrograph of middle saprolite showing the alteration of feldspar to a mixture of halloysite and kaolinite (1), highly altered biotite which has exfoliated and goethite and/or clay minerals between the cleavage planes (2) and opaque mineral (3). (E) Optical micrograph of saprock showing early stages of alteration of feldspar to a mixture of halloysite and kaolinite (1) and partly altered biotite showing a mixture of Fe oxide and clay minerals (2) and quartz (3).

Gibbsite occurs in significant amounts in the bauxite zone and the fragmental duricrust, but decreases appreciably in pisolitic duricrust and pisolitic lag gravels where poorly crystalline alumina (PCA) becomes the major Al mineral

(Figure 5). Gibbsite consists mostly of 0.1-2 µm platy crystals (Anand et al., 1991); some occurs as subhedral to euhedral hexagonal crystals, but most is present as components of basally oriented aggregates. The PCA appears to be a major constituent in near-surface pisoliths and nodules and decreases markedly or is absent in subsurface fragmental duricrust (Anand, 1994). The trend in PCA distribution in the profile is similar to that of hematite and maghemite, but is opposite to goethite and gibbsite. PCA has low loss of ignition (LOI), which is thus an indirect indication of the total gibbsite, goethite and kaolinite content. Two distinct populations for Al_2O_3 versus LOI, one with a positive correlation and one with a negative correlation (Figure 6) suggest that, for a given abundance of Al_2O_3, the LOI is greater in fragmental duricrust and bauxite than pisolitic duricrust and lag, suggesting that most Al in the latter is present as PCA, not gibbsite. Singh and Gilkes (1995) subsequently demonstrated that this unaccounted alumina is χ-alumina, one of several poorly crystalline forms of anhydrous alumina.

Boehmite appears to occur only in small amounts in loose pisoliths and pisolitic duricrust (Grubb, 1971; Davy, 1979; Anand et al., 1991) and, possibly, lower in the profile (Grubb, 1971). Small amounts of relict muscovite occur in duricrust on granitoids. Quartz is abundant (20-25%) over granite, but is much lower (5%) over mafic rocks.

Deeper in the profile, the major change from bauxite to saprolite is marked by the decrease of gibbsite, although in places gibbsite is found in the lower saprolite at close to the weathering front. Halloysite and/or kaolinite are most abundant in the saprolite and are the principal weathering products of feldspars and, to a lesser extent, biotite. Feldspars may weather directly to kaolinite, or via halloysite as an intermediate phase (Anand et al., 1985). Kaolinite crystals typically have platy morphology and are generally <0.5 µm, whereas halloysite crystals are generally tubular but may be partly unrolled or platy especially higher in saprolite profiles. Over mafic rocks, smectite clays are abundant in the saprock and lower saprolite, derived from the weathering of amphiboles and chlorite.

Goethite and hematite generally occur as small platy or sub-rounded crystals in soil and duricrusts. Goethite crystals range from 160-260 µm, whereas hematite ranges from 180-690 µm (Anand and Gilkes, 1987b; Singh and Gilkes, 1992). Where both Fe oxides occurred in the same sample, hematite crystals were always approximately 60% larger than goethite crystals. Aluminium substitution for Fe is quite common in both minerals, affecting the unit cell size, particle size and crystallinity. It is a sensitive indicator of the pedogenic environment of their formation and is dependent on accompanying minerals (Schwertmann, 1988). Substitution of more than 33 mole % Al does not occur in either natural or synthetic crystals (Schwertmann, 1985). Aluminum substitution ranges of 22 to 35 mole % in goethites associated with gibbsite in bauxitic duricrusts in the Darling Range at Boddington, Jarrahdale, Del Park, Huntly and Willowdale (Anand and Gilkes, 1987b; Anand and Paine, 2002). These values are consistent with those for highly weathered materials in Australia and elsewhere (Fitzpatrick and Schwertmann, 1982). Separation of morphologically distinct materials in bauxitic duricrusts in the Darling Range showed that Al substitution in goethite and hematite in the matrix (mean 25 mole %) and pisolith cutans (mean 25 mole%) was systematically lower than for pisolith cores of (29 mole %) (Anand and Gilkes, 1987a). Fitzpatrick and Schwertmann (1982) also found that goethites in ferruginous bauxites from South Africa was highly substituted, containing between 20 and 25 mole % Al and was lower in nodule rinds (14-18 mole %) than cores (20-27 mole %).

There is no significant differences in Al substitution in Fe oxides between lateritic duricrust on granite and dolerite in the Darling Range (Anand and Gilkes, 1987b), despite the different abundances of Fe, Al and other ions. This implies that the amount of Al substitution must reflect formation conditions rather than the gross abundance of ions. In contrast, Singh and Gilkes (1992) found that substitution of Al in goethite was greater for soils on felsic rocks (median value, 26 mole %) than those on alluvial (17 mole %) and mafic (19 mole %) parent materials. However, no reasons were given for these differences.

7 GEOCHEMISTRY OF WEATHERING PROFILES

The behaviour of elements during lateritic weathering in the Darling Range regolith has been discussed by several workers (Sadleir and Gilkes, 1976; Davy, 1979; Davy and El-Ansary, 1986; Ball and Gilkes, 1987; Anand et al., 1991; Anand, 1994). The variation in element abundances in the profile are a function of bedrock geology and the modifying influences of weathering. Each element has a particular pattern of enrichment or depletion through the profile. This can be demonstrated by weathering profiles developed on andesite from Boddington (Figure 7).

The greatest chemical changes between bedrock and the regolith are at the weathering front, the interface between bedrock and saprolite. *Sodium, Ca* and *Mg* are strongly leached from saprolite and the overlying horizons. *Potassium* is also lost from alkali feldspars (Anand and Gilkes, 1987c) but is retained throughout much of the profile in muscovite, which is resistant to weathering. Weathering of biotite to mixed clay minerals and goethite is accompanied by substantial, but not total, leaching of potassium. *Aluminium* is enriched in the weathering profile compared to the bedrock. There is an abrupt rise in Al_2O_3 at weathering front but, above this, there is little variation throughout the overlying clay-rich saprolite and mottled clays until the bauxite zone. *Iron* abundances generally increases upwards through the profile although, on granite, there is depletion from the clay-rich upper saprolite. The abundance of Fe_2O_3 in the duricrust generally reflect the underlying bedrock. The Fe content of the dolerites (12-20% Fe_2O_3) is far greater than that of granite (2-5% Fe_2O_3). This difference is reflected in the duricrusts, which contain 40-45% Fe_2O_3 and 10-12% Fe_2O_3, respectively. Duricrust over andesite contains 25% Fe_2O_3. Other major changes in Fe content occur where residual duricrust on crests and upper slopes passes laterally into transported gravels on lower slopes; however, these changes are not systematic but depend upon the nature of the residual duricrust upslope and on effects related to the transport of Fe (Hickman et al., 1992). *Silica* concentrations are also controlled by the nature of the bedrock, mainly reflecting the abundance of quartz.

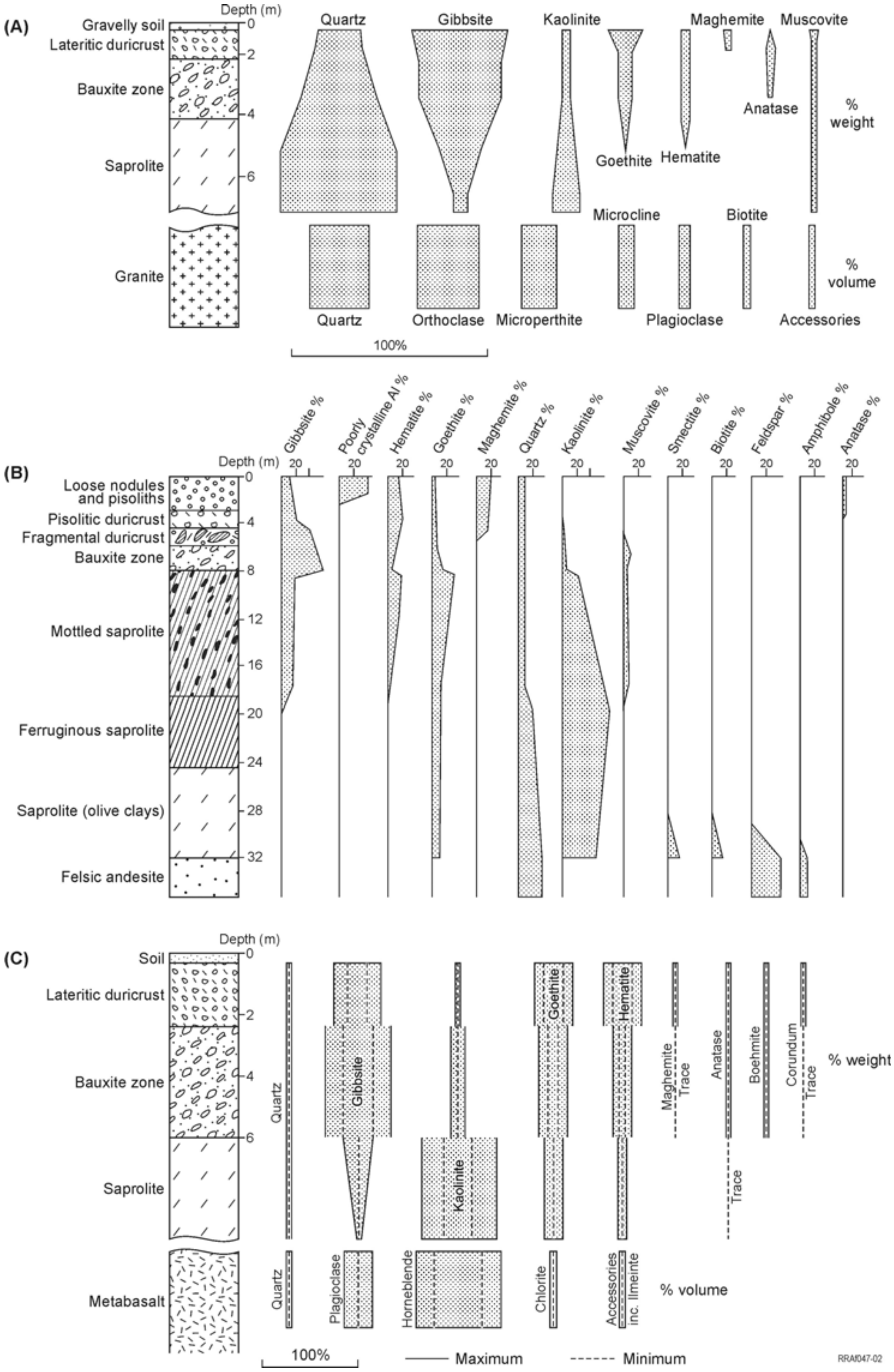

Figure 4: Mineralogy of the Darling Range lateritic bauxite profiles. (A) On granitic bedrock : Jarrahdale (modified from Sadleir and Gilkes (1976). (B) On felsic andesite: Boddington (after Anand, 1994). (C) On metabasalt: Mt Saddleback bauxite deposits (after Ball and Gilkes, 1985).

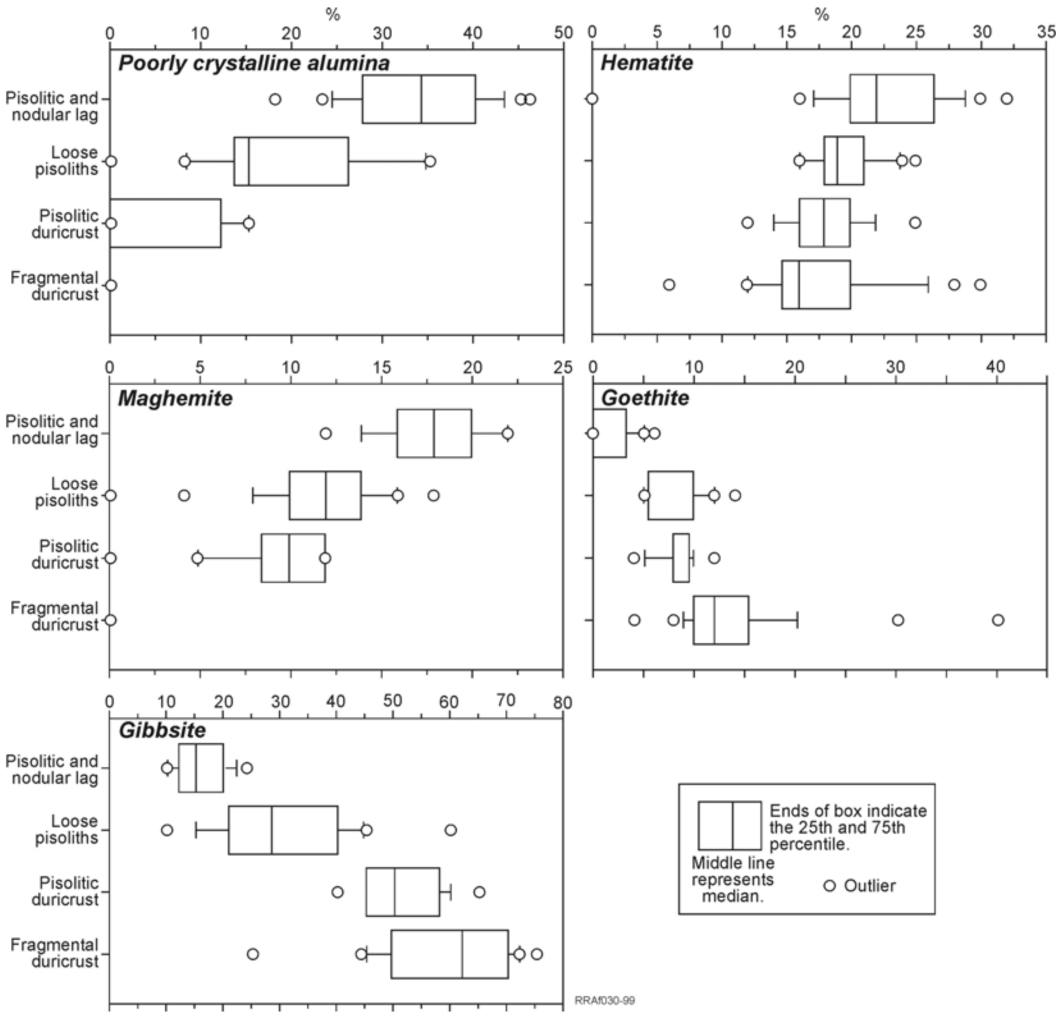

Figure 5: Distribution of poorly crystalline alumina, hematite, maghemite, goethite and gibbsite in the ferruginous zone of bauxitic weathering profiles from the Boddington gold deposits (after Anand, 1994).

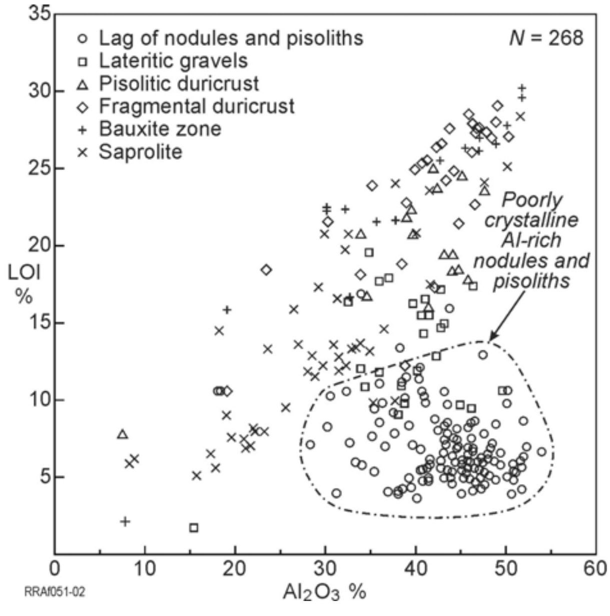

Figure 6. Relationship between Al_2O_3 and loss on ignition (LOI) in regolith units from Boddington gold deposit (after Anand, 1994).

The resistance of quartz to weathering results in its retention throughout the profile into the duricrust, with only small loss by dissolution. Thus, SiO_2 concentration in duricrusts is much greater over the granitoids (15-20% SiO_2) than over dolerite (<5% SiO_2). Total silica contents decrease up profile, with a fairly sharp break between the saprolite and the bauxite zone, due to the alteration of much of the kaolinite to gibbsite.

The distributions of *Ti, Zr, V, Cr, Nb, Th* and *Ga* reflect 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 in zircon; Ti in anatase; Cr in chromite). Their abundances all tend to increase upwards through the profile due to the gradual loss of other components. Chromium, V and Ga tend to accumulate with Fe oxides; Ga is also associated with gibbsite. In consequence, these elements become concentrated in lateritic duricrust, within which lateral dispersion by colluvial action can occur during the course of profile evolution.

Of the trace elements, *Ba, Sr, Cu, Co, Ni* and *Zn* hosted by feldspars and ferromagnesian minerals are commonly leached in the saprolite, although a proportion of Co, Cu and Zn is retained in Fe oxides derived from ferromagnesian minerals. Gold is a significant trace constituent at Boddington, showing moderate concentration in the mid to lower saprolite and in the duricrust, but with strong depletion in the upper saprolite (Davy and El Ansary, 1986).

There may also be some aeolian accession to the regolith. In the Jarrahdale area, Brimhall et al. (1988) distinguished two populations of zircons based on their morphology. Euhedral grains were interpreted as having been derived from the granitic bedrock, whereas rounded grains were regarded as having been derived from aeolian dust and translocated downwards from the surface by illuviation. Foo (1999) determined the concentrations of Zr in different parts of pisoliths at Boddington. He found that Zr is greater in the cutans of pisoliths relative to the cores which may support its derivation in part from external sources during the formation of the cutans.

8 CLASSIFICATION OF LATERITIC DURICRUSTS

8.1 FRAGMENTAL DURICRUST

Fragmental duricrusts generally occur in the lower part of the duricrust. They are composed of angular to subangular, dark red, hematite-gibbsite fragments of various dimensions ranging from a few mm to more than 30 mm, and occasionally 100 mm, across, in a pale gibbsite-rich matrix (Figure 8A). The fragments commonly retain remnant primary lithic fabric with, for example, gibbsite pseudomorphs after feldspars (Figure 8B);voids are infilled with kaolinite and gibbsite. Similar fabrics in duricrust were observed by Hickman et al. (1992) (Figure 9A,B) and outcrops having relict bedrock textures are common (Figure 9B).

8.2 FRAGMENTAL-PISOLITIC DURICRUST

Some fragmental duricrusts consist of large fragments containing hematite and gibbsite in a matrix of gibbsite and goethite that contains pisoliths and ooliths (Figure 8C). In these duricrusts, the fragments are surrounded by red hematitic and goethite cutans that become paler towards the centre. The pisoliths in the matrix are 2-10 mm in diameter, with reddish brown to black, hematite-rich cores surrounded by goethite-rich cutans. The ooliths, in comparison, have cores of goethite, gibbsite and subangular to sub rounded quartz, surrounded by goethitic cutans. The larger ooliths (<2 mm) are themselves set in a matrix of smaller ooliths (50-100 µm), all cemented by goethite (Figure 8D).

8.3 PISOLITIC DURICRUST

The pisolitic duricrusts occur at, or close to, the surface and have a greater variety of fabrics and a more complex mineralogy than the underlying fragmental duricrust. This suggests that individual pisoliths, and different pisolitic duricrusts, have formed under different weathering environments (Figure 8E and F) (Anand, 1994). Two broad types of pisoliths have been identified, black and red. *Black pisoliths* contain minerals that do not occur in the surrounding matrix. They are massive, 5-15 mm in diameter, and are dominated by maghemite, hematite and PCA. The matrix between the pisoliths is yellow to light brown, and consists of gibbsite with minor amounts of goethite and hematite. In contrast, *red pisoliths* are 2-5 mm in diameter, porous and do not contain maghemite; their mineralogy (hematite, goethite and gibbsite) is similar to that of the matrix. Some compound pisoliths occur, having earlier generations of pisoliths within them.

Concentric pisoliths are common. Their nuclei may be massive, composed of either : (a) a mixture of PCA, hematite, maghemite and gibbsite, or (b) hematite and gibbsite; less commonly, the nuclei are lithic fragments, and are composed of gibbsite. Most nuclei, irrespective of their composition, are situated at the centre, or slightly offset from the centre of the pisolith boundaries. Radial desiccation cracks are particularly common within many of the hematite cores, suggesting that hematite resulted from dehydration of goethite. Cutans generally comprise at least 30% of the total pisolith diameter and are mostly concentric, with the innermost cutans having increased thickness so as to fill and smooth out depressions in the nucleus surface. Discordant boundaries (micro-unconformities) occur between cutans, as do thin cutans of silt-sized quartz. Where darker and lighter cutans alternate, the lighter cutans are richer in Al, poorer in Fe and gibbsite-rich.

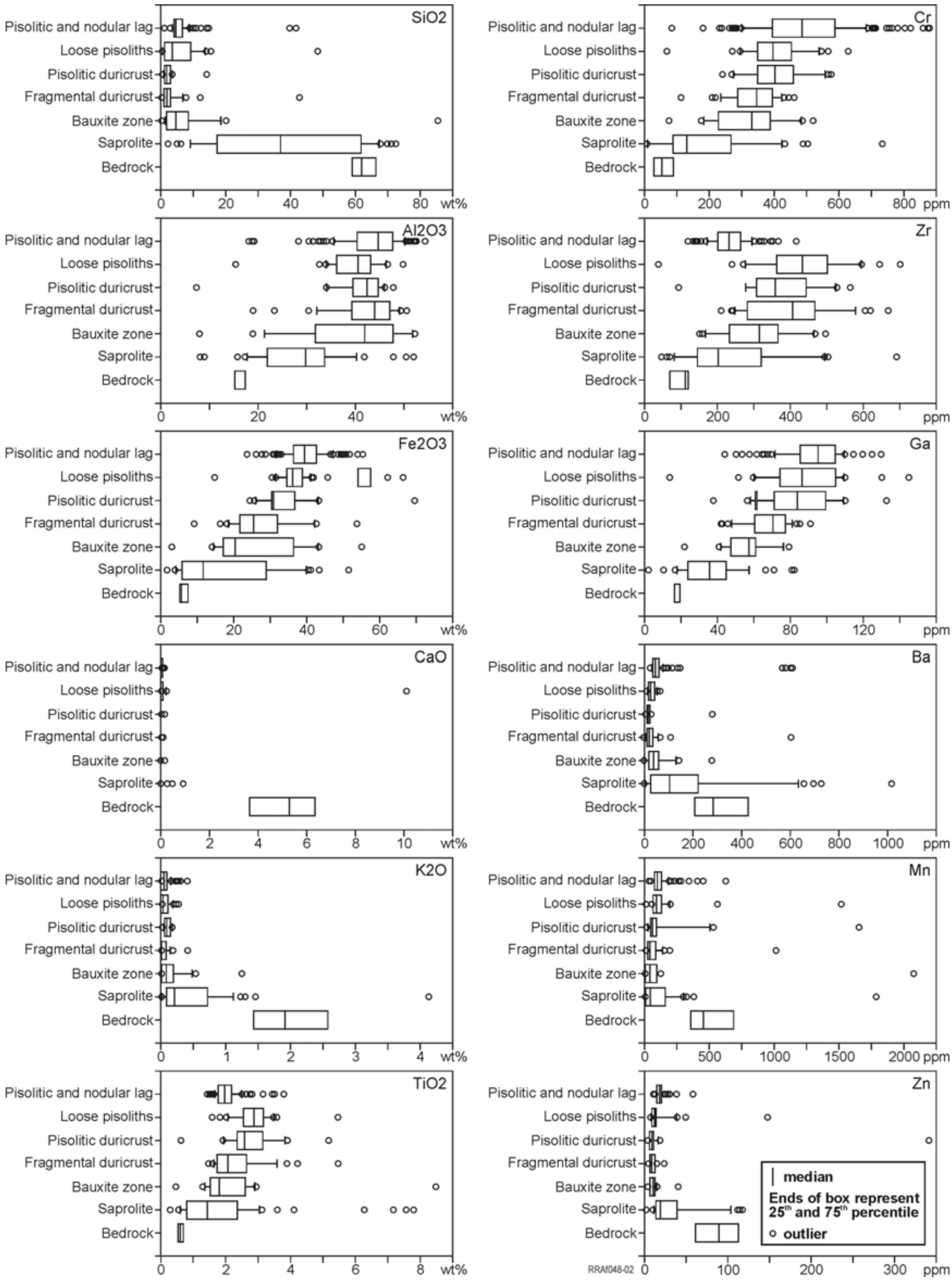

Figure 7: Box plots showing the distribution of major and minor elements in several regolith materials and felsic andesite bedrock arranged vertically in order of the regolith stratigraphy.

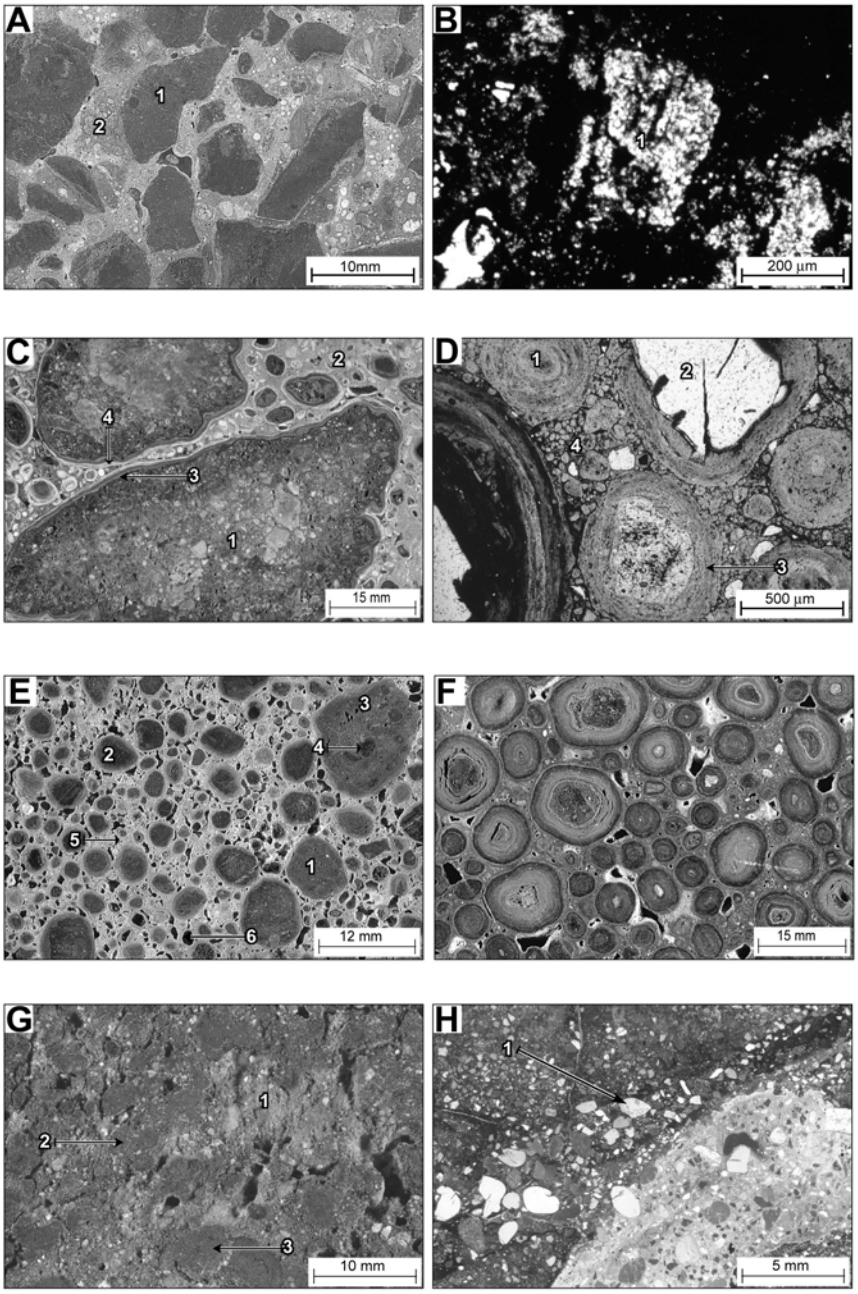

Figure 8: Examples of duricrusts. (A) Fragmental duricrust developed on felsic andesite showing hematite-rich fragments (1) set in a gibbsite-rich matrix (2), Boddington. (B) Photomicrograph of a polished thin section showing gibbsite pseudomophs after feldspars (1). (C) Fragmental-pisolitic duricrust developed from granite showing large, irregular fragments (1) containing hematite, quartz and gibbsite in a matrix of gibbsite and goethite that contains pisoliths (2). The fragments are surrounded by red hematitic (3) and goethitic (4) cutans and become paler towards the centre. (D) Photomicrograph of a polished thin section showing ooliths of different sizes with cores of goethite impregnated gibbsite (1) and sub angular to rounded quartz (2) surrounded by goethitic cuatns (3) set in a matrix of smaller ooliths cemented by goethite (4). (E) Pisolitic duricrust showing pisoliths with cores ranging from earthy red, hematite-stained gibbsite (1) to dark brown to black hematite and maghemite (2) surrounded by red to brown cutans. Some compound pisoliths occur (3) having earlier generations of pisoliths within them (4). The matrix is gibbsitic sandy clay (5) and has numerous small voids (6). (F) Pisolitic duricrust showing concentric pisoliths with a variety of cores. (G) Massive/vesicular duricrust showing sand cemented by goethite (1) and hematite (2). Harder, more hematite-rich areas (3) occur along with angular to rounded quartz grains in both the goethite- and hematite-rich matrix. (H) Photomicrograph of a polished section showing rounded quartz grains (1) embedded in a goethite-rich matrix.

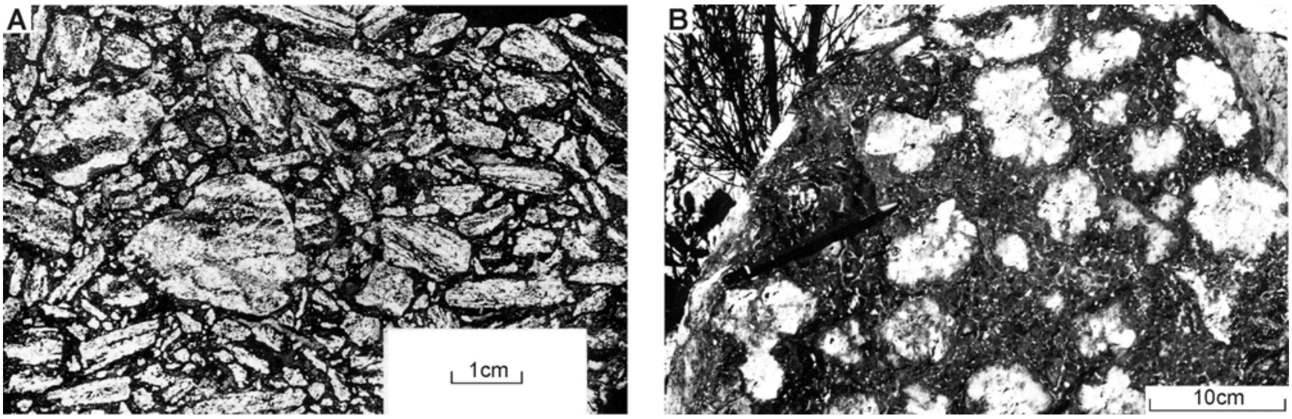

Figure 9: (A) Lateritic duricrust derived from porphyritic granite. Feldspars crystals have been replaced by gibbsite pseudomorphs. Original crystal orientation is preserved (after Hickman et al., 1992). (B) Fabric of porphyritic dolerite preserved in lateritic duricrust (after Hickman et al., 1992).

On lower slopes, some duricrusts have poorly sorted to well sorted pisoliths with yellow goethitic cutans, weakly cemented by outer cutans. Isolated, round frosted quartz grains, reaching 5 mm in size occur as part of the detrital framework. Interstices have resulted in a very porous and permeable duricrust with some sandy yellow clay filling. These duricrusts are formed by erosion of pisoliths from upslope followed by deposition and recementation by goethite on lower slopes.

8.4 VESICULAR DURICRUST

Vesicular duricrusts occur on the valley floors. These duricrusts (Figures 8G and 8H) commonly contain sandy detritus, quartz pebbles and a few ferruginous wood fragments, cemented by goethite. Vesicles (1 to 10 mm) in the goethite matrix vary from rounded to funnel-like or lenticular and irregular. The voids are coated by wart-like protuberances of yellow ochreous goethite. The duricrusts are formed by the precipitation of Fe oxides in seepages and swamps and are equivalent to bog iron ore. Where these are organic, as shown by the presence of ferruginized organic matter, goethite has probably formed from ferrihydrite rather than hematite.

9 DISTRIBUTION OF REGOLITH IN THE LANDSCAPE

Lateritic duricrust occupies gently sloping to horizontal upland areas at 280 to 300 m. Steeper slopes have a thin cover of transported gravels and bedrock is generally near-surface. There is little duricrust or gravel below 200 m. Blocks of duricrust, released by headward erosion of streams, degrade to ferruginous gravel on the lower slopes of valleys. On a local scale, the topography of the weathering front is much more irrregular than that of the top of the duricrust. Pinnacles of the bedrock and isolated corestones occur high in the weathered profile, locally lessening its thickness. Contacts between granite and dolerite can be followed in road cuttings and bauxite pit faces to the surface, despite weathering, indicating that the profile is entirely residual.

Undissected terrain (lateritic upland), characterized by gentle slopes and shallow valleys, has a relatively simple soil distribution. In general, the soil thickness increases with distance from crests towards the valley floor. Sandy gravel is predominant on the slopes and crests, whereas sandy soil increases in thickness towards the valley floor. In the Jarrahdale area, for example, uplands show a toposequence of very gravelly soil (sandy gravel) and lateritic duricrust on the crests to mid slopes (Figure 10). The crest of the ridge may be occupied by large dolerite dykes represented by outcrops of fresh rock enclosed in an aureole of lateritic duricrust. The sandy gravels increase in thickness to more than 1 m on the mid slope and lower slope positions, where they overlie various forms of lateritic duricrust and bauxite. The gravel becomes finer downslope and on lower slopes and is interlayered with sand. ‘Orange earths’ dominate the valley floor but, in some situations, yellow-brown sand may extend on to them. Podzolic soils and cracking clays are also prominent in some valley floors. Here, lateritic duricrust is absent and the soils directly overlie saprolite. Pockets of goethite-rich vesicular duricrust ('bog iron'), formed by chemical or biochemical precipitation of hydromorphically transported Fe, may occur in valley floors. Outcrops of fresh rock are rare.

On long, steep slopes in dissected terrain, the soil pattern is more complex. Rock outcrops are common, and slump features, minor landslides and infilled valley-side depressions indicate a prior more dynamic geomorphic environment than prevails at present. Orange earths dominate the surface with saprolite an important substrate, although fresh rock is common. Locally, convex hills are present, and these have greater thickness of soil.

Local variations in lateritic duricrust morphologies occur within a *catena* or toposequence (i.e. systematic variations down a slope). Fragmental duricrust occurs on crests and upslope positions, whereas pisolitic duricrust becomes abundant on mid-slope positions, overlying fragmental duricrust. In some lower slope positions, pisolitic duricrust may directly overlie saprock or fresh rock. This distribution suggests erosion from crests and upper slopes and deposition on the lower slopes. Vesicular duricrusts occur in valleys. Similarly, bauxite distribution is related to topographic position, being best developed on hill slopes, where leaching is greatest, rather than on less freely drained crests or valley floors.

10 REGOLITH-LANDSCAPE EVOLUTION

10.1 EVOLUTION OF WEATHERING PROFILES

The distribution, and mineralogical and chemical compositions, of the lateritic, mostly bauxitic, regolith of the Darling Range results from the interaction of parent rock, topography, climate, vegetation, drainage and erosion over long periods of time. Weathering has residually enriched Al and Fe, as a result of Si depletion and the almost total loss of Mg, Ca, Na and K. These changes are consistent with dissolution of feldspars, amphiboles, biotite, pyroxene and chlorite, with the retention of Al and some Si in kaolinite and halloysite, of Al in gibbsite and Fe in goethite and hematite. Similarly, other relatively immobile elements (Cr, V, Ga, Ti and Zr) are concentrated in the upper profile, especially the duricrust. However, the boundaries of dolerite dykes, quartz pegmatite veins and other primary features continue without alteration of dips or strike as they pass from bedrock to saprolite and, in places, to the lower duricrust (Baker, 1976, Sadleir and Gilkes, 1976; Davy, 1979). This, together with the preservation of relict fabrics in the saprolite, leads to the conclusion that changes from bedrock to saprolite have been nearly isovolumetric. Where developed, an arenose zone has formed with loss of lithic fabric caused by solution and removal of kaolinite, and settling of quartz. Progressive weathering under well drained conditions results in increased dissolution of silica and a bauxite zone is formed.

The characteristics of pisoliths and nodules in duricrusts suggest two modes of formation - in saprock and soil. Nodules formed in saprock have lithic nuclei that preserve rock fabric. They consist dominantly of gibbsite and hematite; maghemite is typically absent. Fragmental duricrust has resulted from direct gibbsitization of saprock without intermediate formation of a kaolinite-rich deep saprolite. Subsequent ferruginization has helped to protect the rock fabric, except where it is extreme and has progressively destroyed it. Pisoliths formed in soil have homogeneous nuclei, but may have complex, compound structures; they commonly have concentric cutans. They consist largely of hematite, maghemite and PCA. Pisolitic duricrusts with the greatest mineralogical diversity have a far more complex history than mineralogically simpler fragmental duricrusts. Local erosion and deposition would have been continuous throughout the formation of pisoliths, which are cemented by gibbsite and goethite to form pisolitic duricrust. Pisoliths with well developed multiple cutans have a complex, cyclic history and have formed by both accretionary and concretionary processes. Disconformable contacts between cutans, some including thin layers of quartz, indicate that the cutans formed by accretion. Formation of concentric pisoliths by progressive inward hydration to goethite of a pre-existing hematite-rich nucleus, as proposed by Tardy and Nahon (1985), is suggested by irregular dissolution of edges and scattered remnants of the original nucleus within the cutans. However, the latter situation is uncommon.

The pisoliths have formed by replacement or cementation of soil or colluvium by the solution and migration of Fe and the precipitation of Fe oxides in the gibbsite matrix or in voids. This was not a single event, but involved multiple phases of solution and precipitation. The presence of χ-alumina in PCA suggests that aluminium has also been mobilized and reprecipitated. Transitional aluminas, including χ-alumina, are considered to form by the dehydroxylation of hydroxides and oxyhydroxides (Misra, 1986; Wefers and Misra, 1987). This can occur by heating, with the product depending on the rate of heating and the particle size. Thus, dehydroxylation of gibbsite to α–alumina (corundum) under laboratory conditions proceeds via the χ and k forms whereas dehydroxylation of boehmite proceeds via the θ, γ and δ forms (Rooksby, 1961). Heating may have been provided naturally by bush fires. However, based on its morphology, Singh and Gilkes (1995) suggested that the χ-alumina or its precursor was precipitated as a hydrous gel rather than as discrete crystals. The solubility of $Al(OH)_3$ around pH 4 is fairly high (Misra, 1986) so that significant amounts of Al may occur in the strongly desilicated, acid soil solution that exists in the near-surface horizons of bauxitic profiles. In drier seasons, Al may concentrate in soil solution by evaporation and evapo-transpiration, and precipitate as a hydrous amorphous gel in the soil matrix due to increases in pH and concentration (Hsu, 1989). It is possible that organic ions in solution would have inhibited the crystallization of gibbsite or boehmite (Hsu, 1989). Subsequently, this amorphous phase may dehydrate at surface to form transitional alumina and indurate the Fe-rich matrix in successive seasonal depositions.

It is unclear when maghemite formed, although it was probably at a late stage, by heating of goethite and hematite by bush fires (Anand and Gilkes, 1987a). The presence of maghemite thus indicates a very near surface origin, so that the occurrence of magnetic pisoliths at depths greater than 1-2 m implies later burial or incorporation into soil by mechanical processes. The materials above buried magnetic pisoliths are thus transported and unrelated to the underlying regolith or bedrock.

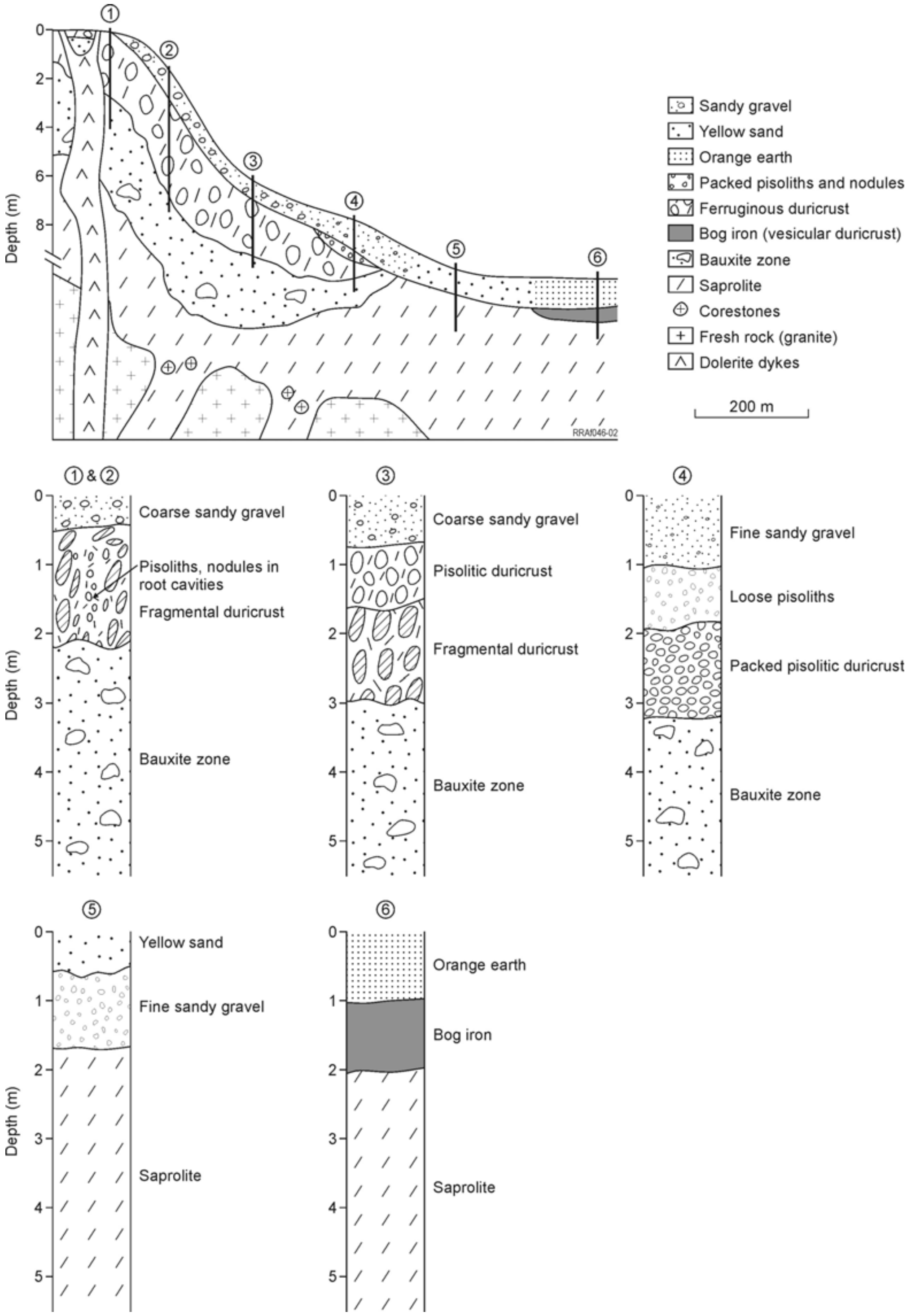

Figure 10: Regolith catena from the Darling Range showing trends in soil and lateritic duricrust following a sequence from crest to valley floor. Sandy gravels on crest and midslope give way to yellow sandy soils on the lower slopes and orange earths in the valleys. The gravels become thicker and finer downslope.

Lateritic duricrusts contain more Fe than the underlying saprolite. Some workers have explained the extra amounts of Fe in relation to the underlying saprolite by general landscape lowering (e.g., Trendall, 1962; Tardy and Roquin, 1992) whereas others (e.g., Ollier and Pain, 1996) by lateral and oblique movement of Fe. Trendall (1962), using the Fe content of granite and the 'laterite' over it in Uganda, calculated that 14 m of granite would be required to produce 0.3 m of 'laterite' and proposed an overall process of ground surface lowering to account for a surface covered with 'laterite'. Up to 3 km of surface lowering has been calculated by Tardy and Roquin (1992) for Madagascar. However, it is not believed that the duricrust in the Darling Ranges is a condensed sequence of this order of magnitude. There appears to have been little or no lowering of land surface, as shown by the preservation of rock fabrics during the formation of saprolitic and fragmental duricrust Dolerite dykes and pegmatite veins exposed in the railway cutting at Jarrahdale can be followed from fresh rock into the duricrust (Sadleir and Gilkes, 1976; Davy, 1979); over the subcrop of the pegmatite, for example, the duricrust contains coarse, angular quartz. In neither case is there a flattening of the dip of the contact with the host rock, which would be expected if the duricrust were part of a condensed profile. Terrill (1956) argued, on similar grounds, that there was no condensed profile over dolerite at Mount Helena and Parkerville. He pointed out that the Al_2O_3/Fe_2O_3 ratio in duricrust over dolerite at Parkerville remained constant from parent rock through to duricrust. Baker (1976) noted that the occurrence of dolerite dykes was clearly identifiable from the Fe_2O_3 content of the bauxite horizon, and that the dolerites could be mapped on the basis of the colour and quartz content of the surface duricrust. However, lowering of a few metres is probable during the formation of pisolitic duricrust. For example, pisolitic duricrust on felsic andesite has 25% Fe_2O_3 and the underlying felsic andesite approximately 6% (Anand, 1994). The concentration factor of 4 is the same for Ti and Zr, from which it can be concluded that the Fe, Ti and Zr, relatively insoluble elements, were concentrated from the weathering of a greater thickness of underlying rock.

The upper parts of the weathering profiles contain variable amounts of aeolian material, but its source is uncertain. It is generally difficult to distinguish exotic aeolian constituents from more locally derived alluvial or colluvial material.

10.2 TOPOGRAPHIC CONTROL AND AGE OF WEATHERING PROFILES

Deep weathering profiles at Jarrahdale and Boddington yield late Tertiary palaeomagnetic ages (Dr B. Pillans, ANU, written communication, July 2002), similar to those reported by Schmidt and Embleton (1976) from the nearby Perth Basin. In contrast, deep weathering profiles in the eastern Yilgarn yield both early and late Tertiary palaeomagnetic ages (Table 16 in Anand and Paine, 2002. The absence of older weathering imprints in the western Yilgarn may well be a consequence of denudation during and since the Mesozoic. The geographical contrast in the depth of post-Permian dissection from west to east is probably the result of differential uplift associated with breakup of Gondwanaland, commencing in the early Cretaceous. Minimal uplift to the east allowed preservation of the relict Permian landscape. There have been various estimates of the amount of material eroded from the Yilgarn Craton (Finkl and Fairbridge, 1979; van de Graaf, 1981; Kohn et al., 1988; Killick, 1998). Finkl and Fairbridge (1979) assumed that there has been little erosion of the Yilgarn Craton since the Proterozoic, whereas van de Graaf (1981) estimated that up to 400 m of erosion has been necessary to generate sediments infilling the surrounding basins. Killick (1998) used a similar mass balance approach to relate the sediment content of the basins surrounding the Western Shield (Pilbara and Yilgarn Cratons, and intervening Proterozoic Basins), and estimated that that as much as 5.2 km may have been eroded from the whole Shield. However, he noted that most of the erosion was Mesozoic or earlier, because Tertiary to Recent sedimentation in the adjacent basins, including the Perth Basin, is dominated by carbonates. Kohn et al., (1998) obtained similar estimates using apatite fission track thermochronology, which suggested erosion of 3-4 km from the area of the Darling Range during the late Palaeozoic and, possibly, an equivalent loss during the Mesozoic. There is strong support for the models of early Phanerozoic denudation along the western margin of the Yilgarn, adjacent to the Perth Basin. The geomorphological and sedimentary record of Late Palaeozoic glaciation appears to have been completely removed, with the exception of small outliers of Permian glacial strata and coals preserved in depositional rifts (Mishra, 1996). It should be noted, however, that extensive denudation models, such as that of van de Graaf (1981), have been questioned by Sircombe and Freeman (1999) and Cawood and Nemchin (2000) who identified low concentrations of shield-derived zircon grains in deposits of the Perth Basin. They suggested that this is evidence that the western part of the shield was not a significant source of sediments during the Phanerozoic and, thus, unlikely to have experienced extensive stripping.

There has been a great deal of discussion on the role of topography on the formation of bauxitic weathering profiles. Hickman et al. (1992) summarize many authors' views on the evolution of bauxite in relation to topography, without committing to a preferred interpretation. Woolnough (1918) suggested that near sea level peneplain conditions were required for 'laterite' formation and that its present topographic high positions resulted from subsequent uplift. Playford (1959) suggested that during lateritization the land surface must have displayed more relief than remains today because a watertable always shows less relief than its overlying surface. By contrast, Terrill (1948, 1956) implies that the topography at the time of lateritization closely resembled that seen today. The topography of the bauxitic deposits is controlled by the nature of the bedrock; faults define stream courses, thicker dolerites underlie ridges (Davy, 1979;

Hickman et al., 1992). If the bauxite formed in the past on a peneplain, which was then uplifted and subsequently eroded, the thickest bauxite would not now lie on the mid slopes. If they formed in the past on topography similar to that of today, it follows that all the erosion that presumably followed the rejuvenation of the streams did not modify the established bauxite-topography relationships. It appears that the bauxites in the Darling Range are forming today (Bettenay et al., 1976, Anand and Paine, 2002). This is consistent with the decrease in bauxitic grade with the present rainfall, i.e. in an easterly direction, and with the work of Dammer et al. (1999) who found the zero age for cryptomelane from Mundijong and suggested that, at present, intense weathering might still be active in areas with abundant precipitation (1000 mm/y and more). The dominance of older ages of K-Mn oxides (between 36 and 20 Ma) obtained in the central part of the Yilgarn Craton, relative to those obtained from southern and southwestern coastal areas, may reflect the time when climate in the Eastern Goldfields became too arid to enable the formation of K-Mn oxides in the regolith. This may indicate that differences in climate similar to present conditions between the central and coastal regions have existed since the late Miocene (Dammer et al., 1999). The seasonal climate provides a favourable environment for the strong leaching and active formation of gibbsite under the winter rainfall of the present Mediterranean climate (Anand et al., 1985), compared to lesser leaching and the formation of kaolinite in uplifted plateaux in savannas having a similar rainfall during the summer (Butt and Zeegers, 1992).

11 REFERENCES

Anand, R.R., Gilkes, R.J., Armitage, T.M. and Hillyer, J.W., 1985. Feldspar weathering in lateritic saprolite. *Clays and Clay Minerals* 33: 31-43.

Anand, R.R. and Gilkes, R.J., 1987a. The association of maghemite and corundum in Darling Range laterites, Western Australia. *Australian Journal of Soil Research* 35: 303-11.

Anand, R.R. and Gilkes, R.J., 1987b. Iron oxide in lateritic soils from Western Australia. *Journal of Soil Science* 38: 607-22.

Anand, R.R. and Gilkes, R.J., 1987c. Muscovite in Darling Range laterites. *Australian Journal of Soil Research* 25: 445-50.

Anand, R.R., Gilkes, R.J. and Roach, G.I.D., 1991. Geochemical and mineralogical characteristics of bauxites, Darling Range, Western Australia. *Applied Geochemistry* 6: 233-248.

Anand, R.R., 1994. Regolith-landform evolution and geochemical dispersion from the Boddington gold deposit, Western Australia. CSIRO Division of Exploration and Mining, Restricted Report 24R, 156pp. (Reissued as CRCLEME Open File Report 3, Perth, 1998).

Anand, R.R. and Paine, M., 2002. Regolith geology of the Yilgarn Craton, Western Australia: implications for exploration. *Australian Journal of Earth Sciences* 49: 3-162.

Baker, G.F.U., 1972. Origin of Darling Range bauxites, Western Australia. *Economic Geology* 60; 215-225.

Baker, G.F.U., 1976. Bauxite deposits of the Darling Range, Western Australia. In: C.L. Knight (Editor), The Economic Geology of Australia and Papua-New Guinea. Australasian Institute of Mining and Metallurgy, Melbourne.

Ball, P.J. and Gilkes, R.J., 1987. The Mt Saddleback bauxite deposit, South Western Australia. *Chemical Geology* 60 215-225.

Brimhall, G.H., Lewis, C.J., Ague, J.J., Dietrich, W.E., Hampel, J., Teague, T. and Rix, P., 1988. Metal enrichment in bauxites by deposition of chemically mature aeolian dust. *Nature* 333: 819-24.

Butt, C.R.M., 1981. The nature and origin of the weathering mantle, with particular reference to Western Australia. In: H.A. Doyle, J.E. Glover and D.I. Groves (Editors), *Geophysical Prospecting in Deeply Weathered Terrains*. Publications Geology Department and Extension Service, University of Western Australia, 6 (1981) 11-29.

Butt, C.R.M. and Zeegers, H., (Editors), 1992. Regolith exploration geochemistry in tropical and subtropical terrains. Handbook of Exploration Geochemistry 4. Elsevier, Amsterdam, 604pp.

Calvert, C.S., Buol, S.W. and Weed, S.B., 1980. Mineralogical transformations of a vertical rock-saprolite-soil sequence in the North Caroline Piedmont. *Soil Science Society of America Journal* 44:1096-1112.

Cawood, P.A. and Nemchin, A.A., 2000. Provenance record of a rift basin: U/Pb ages of detrital zircons from the Perth Basin, Western Australia. *Sedimentary Geology* 134: 209-234.

Churchman, G.J. and Gilkes, R.J., 1989. Recognition of intermediates in the possible transformation of halloysite to kaolinite. *Clays Minerals* 24: 579-590.

Churchman, G.J., 1999. The alteration and formation of soil minerals by weathering. In: M. E. Summer (Editor), Handbook of Soil Science, CRC Press, Boca Raton, Fla., USA. F3-F76.

Churchward, H.M. and McArthur, W.M., 1980. Landforms and soils of the Darling System, Western Australia: In: Atlas of Natural Resources, Darling System, Western Australia: Department of Conservation and Environment, pp.25-33.

Coventry, R.J., Taylor, R.M. and Fitzpatrick, R.W., 1983. Pedological significance of the gravels in some red and grey earths of central north Queensland. *Australian Journal of Soil Research* 21: 219-40.

Dammer, D., McDougall, I. and Chivas, A.R., 1999. Timing of weathering-induced alteration of manganese deposits in Western Australia: evidence from K/Ar and 40Ar/39Ar dating. *Economic Geology* 94: 87-108.

Davy, R., 1979. A study of laterite profiles in relation to bedrock in the Darling Range, near Perth, W.A. Geological Survey of Western Australia Report. 8, 87 pp.

Davy, R. and El-Ansary, M., 1986. Geochemical patterns in the laterite profile at the Boddington gold deposit, Western Australia. *Journal of Geochemical Exploration* 26: 119-44.

Eswaran, H., and Wong, C.B., 1978. A study of a deep weathering profile on granite in Peninsular Malaysia. Parts I, II and III. *Soil Science Society American Journal* 42: 144-158.

Finkl, C.W. and Fairbridge, R.W., 1979. Paleogeographic evolution of a rifted cratonic margin: S.W. Australia. *Palaeogeography, Palaeoclimatology, Palaeoecology* 26: 221-52.

Fitzpatrick, R.W. and Schwertmann, U., 1982. Al-substituted geothite – an indicator of pedogenic and other weathering environments in South Africa. *Geoderma* 27:335-347.

Foo, M.F.P., 1999. Host of gold and ore related elements in ferruginous materials in mineralised environments - implications for exploration.W.A. Honours Thesis, School of Applied Geology, Curtin University of Technology, Perth, 133pp. (Unpublished).

Glassford, D.K. and Semenuik, V., 1995. Desert-aeolian origin of Late Cainozoic regolith in arid and semi-arid South-western Australia. *Palaeogeography, Palaeoclimatology, Palaeoecology* 114: 131-66.

Grubb, P.L.C., 1971. Mineralogical anomalies in the Darling Range bauxites at Jarrahdale, Western Australia. *Economic Geology* 66: 1005-16.

Hickman, A.H., Smurthwaite, A.J., Brown, I.M. and Davy, R., 1992. Bauxite mineralisation in the Darling Range, Western Australia. Geological Survey of Western Australia Report 33, 83pp.

Hsu, P. A., 1989. Aluminium hydroxides and oxyhydroxides. In: Minerals in Soil Environments (J.B. Dixon and S.B. Weed, Editors) Soil Science Society America, Madison, Wisconsin, USA, pp. 331-378.

Killick, M.J., 1998 Phanerozoic denudation of the Western Shield, Western Australia. *Geological Society of Australia Abstracts* 41: 248.

Kohn, B.L., O'Sullivan, P.B., Gleadow, A.J.W., Karner, G.D. and Weissel, J.K., 1998. Late Palaeozoic cooling of southwest Australian terranes inferred from apatite fission track thermochronology. *Geological Society of Australia Abstracts* 41: 253.

McFarlane, M.J., 1976. Laterite and Landscape. Academic Press, London, 151 pp.

Milnes, A.R., Bourman, R.P. and Northcote, K.M., 1985. Field relationships of ferricretes and weathered zones in southern South Australia: a contribution to `laterite' studies in Australia. *Australian Journal of Soil Research* 23: 441-65.

Misra, C., 1986. Industrial Alumina Chemicals. ACS monograph, 184.

Mishra, H.K., 1996. Comparative petrological analysis between the Permian coals of India and Western Australia: palaeoenvironments and thermal history. *Palaeogeography, Palaeoclimatology, Palaeoecology* 125: 199-216.

Mulcahy, M.J., Churchward, H.M. and Dimmockj, G.M., 1972. Landforms and soils on an ullifted peneplain in the Darling Range, Western Australia. *Australian Journal of Soil Research* 10: 1-14.

Murray, A.M., 1979. Bauxite. In: R.T. Prider (Editor) Mining in Western Australia, University of Western Australia Press, Perth, pp. 101-110.

Ollier, C.D., and Pain, C.F., 1996. Regolith, Soils and Landforms. John Wiley and Sons, Chichester, 316pp.

Playford, P.E., 1959. Jurassic stratigraphy of the Geraldson district, Western Australia. *Journal of Royal Society of Western Australia* 42:101-124.

Robertson, I.D.M. and Eggleton, R.A., 1991. Weathering of granitic muscovite to kaolinte and halloysite and of plagioclase-derived kaolinite to haolloysite. *Clays and Clay Minerals* 39:113-126.

Sadleir, S.B. and Gilkes, R.J., 1976. Development of bauxite in relation to parent material near Jarrahdale, Western Australia. *Journal of the Geological Society of Australia* 23: 333-44.

Schmidt, P.W. and Embleton, B.J.J., 1976. Palaeomagnetic results from sediments of the Perth Basin, Western Australia, and their bearing on the timing of regional lateritisation. *Palaeogeography, Palaeoclimatology, Palaeoecology* 19: 257-273.

Schwertmann, U. and Fetcher, H., 1984. The influence of aluminium on iron oxides. XI. Aluminium substituted maghemite in soils and its formation. *Soil Science Society American Journal*. 48:1462-1463.

Schwertmann, U., 1985. The effect of pedogenic environments on iron oxide minerals. Advances in Soil Science 1:171-200. Springer-Verlag, New York.

Schwertmann, U., 1988. Occurence and formation of iron oxides in various pedoenvironments. In: J. W. Stucki, B. A. Goodman and U. Schwertmann (Editors), Iron in Soils and Clay Minerals. Dordrecht: D. Reidel Publishing Company, pp. 267-308.

Schwetmann, U. and Taylor, R.M., 1989. Iron Oxides. In Dixon, J.B. and S.B. Weed (Editors) Minerals in Soil environments pp. 379-438. Soil Science Society of America, Madison, WI.

Singh, Balbir and Gilkes, R. J., 1992. An electron-optical investigation of the alteration of kaolinite to halloysite. *Clays and Clay Minerals* 40, 212-229.

Singh, Balbir. and Gilkes, R.J., 1995. The natural occurrence of x-alumina in lateritic pisolites. *Clay Minerals* 30: 39-44.

Singh, B. and Gilkes, R.J., 1992. Properties and distribution of Fe oxides and their association with minor elements in the soils of south-western Australia. *Journal of Soil Science* 43: 77-98.

Sircombe, K.N. and Freeman, M.J., 1999. Provenance of detrital zircons on the Western Australia coastline-implications for the geologic history of the Perth Basin and denudation of the Yilgarn Craton. *Geology* 27: 879-882.

Tardy, Y. and Nahon, D., 1985. Geochemistry of laterites, stability of Al-goethite, Al-hematite and Fe3+-kaolinite in bauxites and ferricretes: an approach to the mechanism of concretion formation. *American Journal of Science* 285: 865-903.

Tardy, Y. and Roquin, C., 1992. Geochemistry and evolution of lateritic landscapes. In: I. P. Martini, and W. Chesworth (Editors), Weathering, Soils and Paleosols. Amsterdam: Elsevier, pp. 407-43.

Trendall, A.F., 1962. The formation of apparent peneplains by a process of combined lateritisation and surface wash. *Zeitschrift Fur Geomorphologie (Annals of Geomorphology)* 6: 183-197.

Taylor, R.M.and Schwertmann, U., 1974. Maghemite in soils and its origin, Parts 1 and 11. *Clays and Clay Minerals* 10: 289-310.

Terrill, S.E., 1956. Laterite and materials of similar appearance in South-Western Australia; Presidential Address, July, 1954: *Journal of the Royal Society of Western Australia* 40: 1-4.

van de Graaff, W.J.E., 1981. Palaeogeographic evolution of a rifted cratonic margin: S.W. Australia, a discussion. *Palaeogeographic, Palaeoclimatology, Palaeoecology* 34: 163-72.

Wefers, K. and Misra, C., 1987. Oxides and Hydroxides of Aluminium. Alcoa Technical Centre USA.

Woolnough, W.G., 1918. The physiographic significance of laterite in Western Australia. *Geological Magazine* 5: 385-93.