

6 Rock weathering and structure of the regolith

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6.1 INTRODUCTION

The previous two chapters have considered the mineralogical and chemical changes that occur during weathering. As rocks weather chemically, and their mineral constituents change to new, more stable assemblages, the contained elements are preserved in resistate minerals, partly redistributed into new minerals or taken into solution – in some cases to be incorporated in other parts of the weathering profile. Such elemental changes within the weathering profile may be large and, in some cases, result in economic mineral deposits (for example, the concentration of Al in bauxite deposits; Chapter 1). This chapter considers the mineralogical and geochemical changes that accompany progressive rock weathering and discusses the typical weathering profiles for common rock compositions.

6.2 STRUCTURE OF THE REGOLITH

The structure of the regolith at any particular site depends on the extent to which chemical weathering has transformed the bedrock composition, as well as the degree of physical and chemical addition and removal of materials. Well-developed profiles show a

vertical zonation, which may include from depth to surface:

- a zone of partially weathered bedrock that retains the primary rock fabric
- a clay-rich or sandy plasmic/arenose zone in which the primary rock fabric has been destroyed
- a ferruginous mottled zone; a ferruginous, bauxitic or siliceous duricrust/residuum
- a soil layer
- a surface lag of chemically and physically resistant materials (Figure 6.1).

The zone in which the primary rock fabric is preserved is referred to as the *saprolith*. The zone in which the parent fabric has been destroyed, new fabrics formed or soil developed is termed the *pedolith*. Weathering occurs throughout the profile down to the *weathering front*, which is defined as the boundary between fresh rock and saprolith (that is, rock that shows some sign of chemical weathering). Depending on bedrock type and landscape setting, various parts of this mature zonation may be absent, eroded or buried. Across a landscape (or paleolandscape), there is generally significant lateral variation in the regolith and its chemical structure, and these variations may be down to a

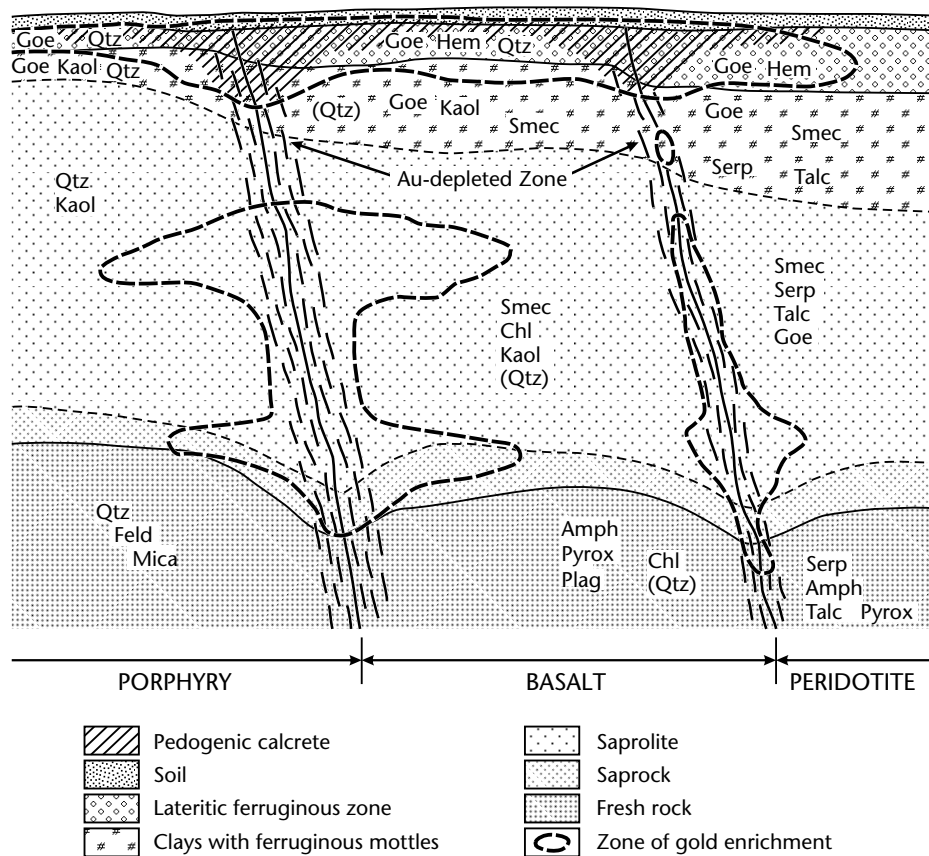


Figure 6.2: Supergene enrichment in Au caused by concentration of Au derived from higher in the weathering profile (after Butt 1989).

highly weatherable pyroxenes and apatite versus resistate zircon) and the weathering conditions. The light REE – particularly Ce, (probably as Ce^{4+}) – are mobile under strongly oxidising conditions (see Section 6.8 below).

Major primary mineral alteration, chemical leaching and secondary mineral growth can eventually destroy the primary rock fabric to produce the pedolith (Figure 6.1). This typically contains a clay- or quartz-rich plasmic/arenose zone (strongly enriched in Al and Si relative to the parent bedrock) and a mottled zone (in which darker Fe oxides are segregated from the more pallid clay minerals). Under some conditions, weathering in the plasmic zone can result in the alteration of kaolinite to gibbsite and leaching of Si. Most primary quartz and resistate accessory minerals are retained – resulting in residual enrichment in Si and elements such as B, Cr, Hf, Nb,

Rb, REE, Th, Ti, V, W and Zr. The mottled zone in the upper part of the pedolith with Fe- and clay-rich zones is enriched in Fe relative to the zone below. Marked accumulation of the Fe oxides – particularly over mafic and ultramafic rocks – can produce a very ferruginous zone, which is typically composed of ferricrete (if cemented by Fe oxides) or ferruginous residuum (if less consolidated). This very ferruginous zone is enriched in elements associated with hematite and goethite, as well as those present in highly resistate accessory minerals. Accumulation of Al oxides, such as gibbsite and boehmite, can form a bauxitic zone and precipitation of dissolved silica can produce an almost pure SiO_2 accumulation of silcrete. These chemically stable and, in most cases, physically resistant materials can harden to form a duricrust and protect this zone of the regolith from erosion and further major chemical alteration.

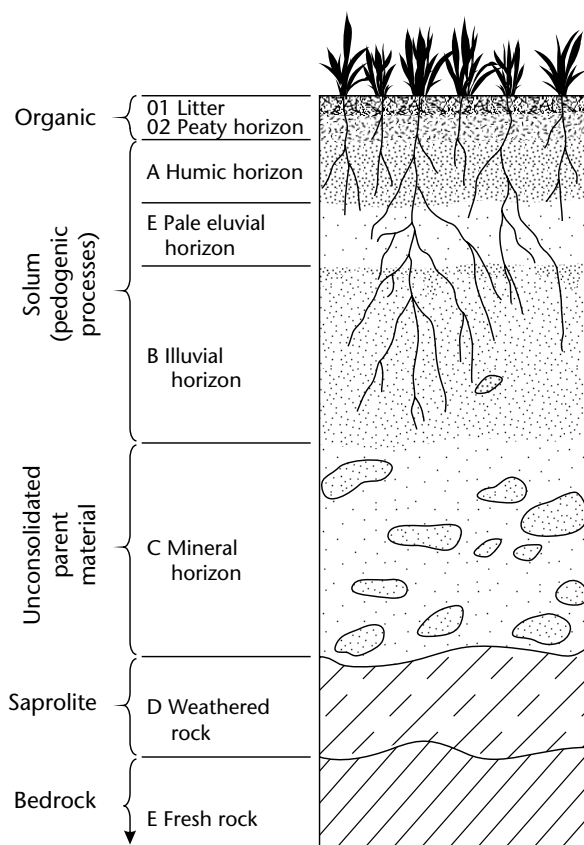


Figure 6.3: An idealised soil profile (after Butt *et al.* 2005).

A *soil profile* will generally develop close to the surface in the zone of greatest biological activity and accumulation of organic detritus. Decay of organic matter and the activity of microorganisms and vegetation result in concentration of humic acids, organic complexes, carbonates and nitrates. The soil chemistry will reflect climate and water content as well as the composition of the underlying regolith/bedrock and any colluvial, alluvial and aeolian additions. See Section 13.2.2 for more detail on soils as a sampling medium.

A mature soil in equilibrium with the local environment is typically differentiated into a number of *horizons*. These are soil layers that are approximately parallel to the land surface and differ from adjacent, genetically related, layers in their physical, chemical and/or biological properties, or in characteristics such as colour, pH, structure, fabric, texture, consistency, and types and number of organisms. Soils may vary in

character from thin, coarse-grained lithosols – that have only one or two poorly differentiated horizons – in some deserts, to thick, organic-, silt- and/or clay-rich soils differentiated into several horizons, in more humid regions. Common soil horizons (Figure 6.3) include:

- Litter (or O1): organic matter on the ground surface
- O (or O2): fibrous (peaty) or massive organic matter
- A: near-surface mineral horizon containing humified organic matter
- E (or A2): pale, commonly sandy, eluvial horizon with little organic matter. Iron and Mn oxides and clays leached or translocated to lower horizons
- B: illuvial horizons enriched in clay, and/or Fe and Mn oxides and/or organic matter derived from overlying horizons
- C: mineral horizon from which the overlying horizons are presumed to have been derived. It is only slightly affected by pedogenic processes so that remnant geological structures or fabric may be retained
- D: layers below the C horizon unaffected by the pedogenic processes that formed overlying horizons, such as previously formed saprolite or transported overburden
- R: continuous fresh rock.

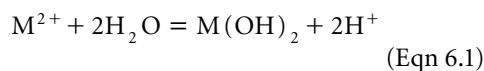
The A, E and B horizons are referred to as the *solum*.

This terminology is ideally suited to well-differentiated soils, such as podzols, but less readily applied to arid-zone soils that have weak horizon differentiation. In soils in semi-arid Australia, the B horizon may be differentiated from the A horizon by a contrast in texture and/or being more sodic or alkaline, rather than the characteristics noted above. For more detail on soil types see CSIRO (1983) for Australia and FAO-UNESCO (1988) for a more global view.

A *surface lag* may form above the soil if the finer or less dense material is preferentially removed by sheet-wash or wind erosion. As quartz and hematite are the most chemically and physically stable minerals under most surface conditions, this lag is relatively enriched in Fe and Si. The presence of stable accessory minerals

in the lag or underlying duricrust can also result in relative enrichment in Ba, Cr, Mn, Nb, Ti, W and Zr. Other minor and trace elements can be accumulated and concentrated in the Fe and Mn oxides that persist in the lag (such as As, Bi, Co, Cr, Ga, Ni, Pb, Sc, Th and V).

The *water table* is an important chemical and physical boundary within the weathering profile – marking the interface between the zone of water saturation and the overlying zone of partial water content. In weathering profiles that are in a steady state, the water table will generally occur in the saprock or lower saprolite zone. However, its position can change seasonally, or over longer time periods, with climate change and landscape evolution. Under wetter conditions the water table can rise up through the weathering profile. As conditions become drier, the water table may progressively fall. The depth to the water table will also depend on the geomorphic setting and local hydrologic factors, including vegetation pumping. Below the water table, conditions are typically reducing (low Eh), and water movement is generally slow through the water-filled fractures and pore spaces (Figure 6.4). Permeability and hydraulic gradient control water movement. Above the water table, gases (including oxygen from the atmosphere) have greater access to the regolith via voids and fractures and conditions are typically more oxidising (higher Eh). Water generally moves rapidly through this zone transporting dissolved and suspended components and leading to strong leaching with further void formation. As well as moving downwards and laterally, water and dissolved components can also move upwards through this zone by capillary action – particularly where evaporation exceeds rainfall. The change in Eh near the water table is commonly sufficient to produce a redox zone marked by the precipitation of insoluble Fe and Mn oxides. The pH may also be reduced in this zone due to the release of H^+ when divalent cations dissolve:



Acid production is even greater if the cation is Fe^{2+} , which may be oxidised and hydrolysed (sometimes referred to as *ferrolysis*):

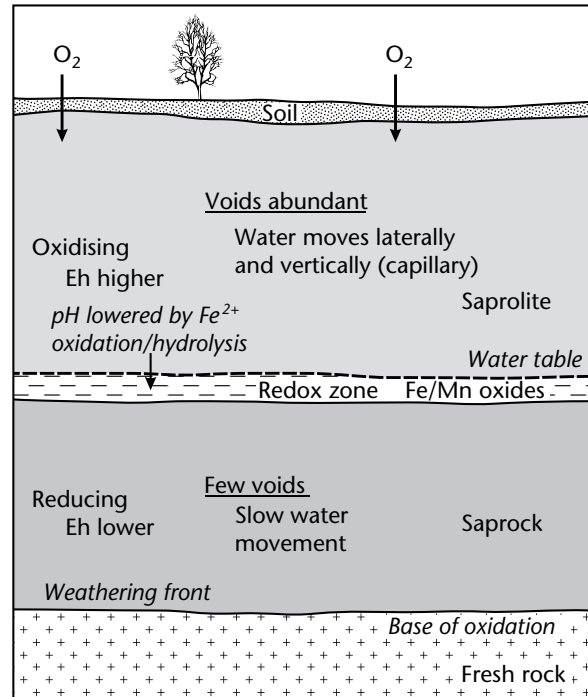
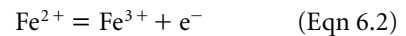
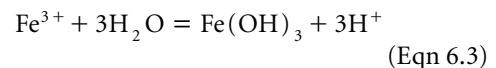


Figure 6.4: Eh and pH conditions about the water table.



and



Thus acidity is generated during chemical weathering even in the absence of sulfides (see Chapter 5.3.1 eqns 5.22–5.25 and Chapter 10).

Groundwater compositions can also have a significant affect on the chemical conditions above and below the water table, including on a regional scale. For example, in the southern Yilgarn Craton of Western Australia, conditions in the upper 5–30 m of the regolith are commonly acid and oxidising, whereas in the northern Yilgarn Craton they are commonly neutral and weakly to moderately oxidising (Gray 2001).

6.3 FIELD EXAMPLES OF WEATHERING PROFILES ON COMMON ROCK TYPES

Particular zones within a regolith profile are commonly recognised by their colour, induration, textural

features and mineralogy (such as recognition of mottling). Some of these features can be difficult to determine in pulverised and homogenised drill chips – a common means of observing and sampling weathering profiles during mineral exploration. Field descriptions of zones within profiles based on information obtainable from drill chips are therefore simplified from those in Figure 6.1. Thus, in the descriptions of weathering in common rock types (below), the weathered materials are divided into lower saprolite (which also includes saprock) and leached upper saprolite (which may include some pedolith material), respectively. Examples are for complete, well-developed and largely intact profiles. Such profiles are more commonly preserved in areas with a long history of weathering, tectonic stability and low relief or where they are protected from erosion by well-developed duricrusts or other deposits (such as lava flows). As well as climatic conditions and geomorphic setting, the degree and depth of weathering are influenced by rock type, structure and the presence of sulfide mineralisation. Studies in the Yilgarn Craton of Western Australia indicate that the extent and depth of weathering commonly follow the succession: granite and mafic rocks < ultramafic rocks < sediments and fine-grained felsic rocks (Anand and Paine 2002). Over areas of strong fracturing and sulfide mineralisation, the depth of weathering can be considerably greater.

6.3.1 Weathering in granitoids

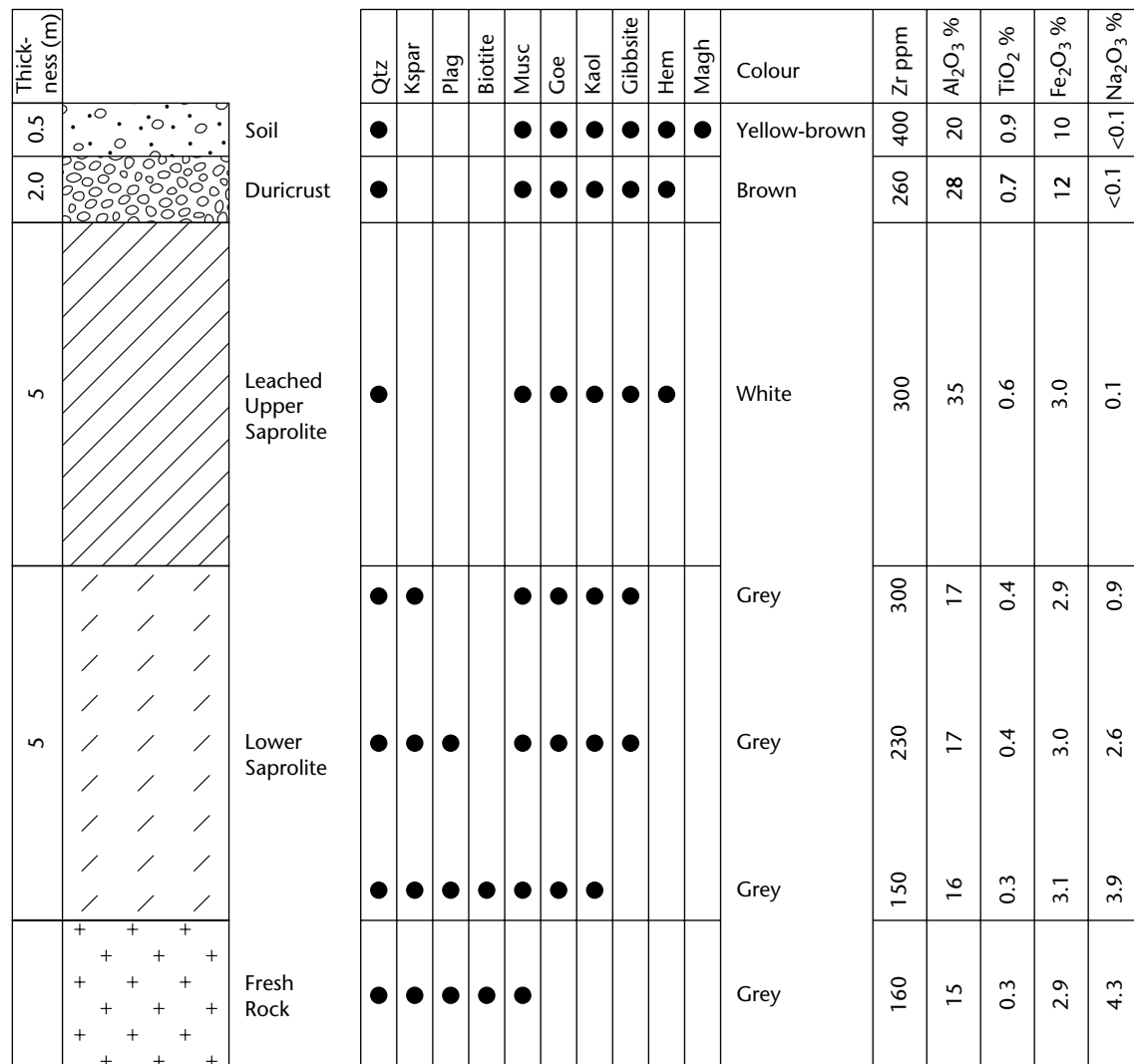
Granitoids typically have a massive fabric, and weathering is generally initiated by water penetration along regular unloading joints and other fractures. The major minerals in unweathered granitoids are quartz, K-feldspar, plagioclase, biotite and, in some cases, muscovite and amphibole. With the commencement of weathering, the ferromagnesian minerals (biotite and amphibole) and feldspars start to breakdown to form goethite and kaolinite so that Al and Fe are retained but more soluble components (Mg, Na, K and Ca) are lost from the lower saprolite. With more intense/longer weathering these processes continue so that feldspars and ferromagnesian minerals are completely destroyed and kaolinite becomes a major component of the leached upper saprolite – resulting in relative enrichment of Al (Aspandiar 1998). Minor

amounts of other clays, such as halloysite and smectite, may also form. Further weathering – particularly under humid climatic conditions –, can lead to breakdown of kaolinite and removal of silica in solution to produce Al hydroxides (gibbsite and boehmite): that is, a bauxite zone above the kaolinite-rich saprolite. Some elements that are retained in resistate minerals – such as Zr and Ti, in zircon and rutile – are residually concentrated in the upper part of the profile (Butt 1985; Figure 6.5). Soluble Na, K and Ca are strongly leached from the upper saprolite. The K present in muscovite is, however, retained. In the example of granitic weathering from the Darling Ranges, WA (Figure 6.5), the formation of a surficial siliceous capping (duricrust) has diluted all other components, but this horizon, and the thin soil formed above it, contain the same minerals as in the leached upper saprolite – though not in the same proportions.

Weathering profiles in felsic volcanic rocks show a similar pattern of mineral development to weathering in granitoids, but with stronger ferruginous mottling where they have a higher original Fe content (Anand and Paine 2002).

6.3.2 Weathering in mafic rocks

Most mafic rocks are igneous or metamorphosed igneous rocks and generally contain dominant plagioclase, ferromagnesian minerals (such as olivine, pyroxenes, amphiboles, biotite and chlorite) and variable amounts of other minerals, including quartz. Muscovite, carbonates and sulfides (mainly pyrite or pyrrhotite) may be present in alteration and mineralised zones. With the commencement of weathering, the carbonate minerals (dolomite and calcite) rapidly dissolve, pyrite and pyrrhotite weather to form goethite, and the ferromagnesian minerals alter to smectite and kaolinite. This results in loss of Ca, but retention of Al, Fe and Mg (the latter two may even be slightly residually enriched), at the base of the lower saprolite. With more intense/longer weathering, plagioclase and chlorite also break down to form smectites and more kaolinite and goethite, with Mg being lost from higher in the lower saprolite. Further weathering results in the complete loss of plagioclase and chlorite and the formation of more clay minerals (kaolinite and smectites) and goethite/hematite, with



Key to Figures 6.5 - 6.9

Calc Calcite
 Chl/V Chlorite/vermiculite
 Dol Dolomite
 Goe Goethite
 Gyp/Ba Gypsum/barite

Hem Hematite
 Kaol Kaolinite
 Kspar K-feldspar
 Magh Maghemite
 Musc Muscovite

Parag Paragonite
 Plag Plagioclase
 Qtz Quartz
 Smec Smectite

Figure 6.5: A generalised profile through weathered granite, Darling Ranges, WA (after Gilkes *et al.* 1973; Anand 1984).

soluble elements – particularly K and Mg – being lost from the leached upper saprolite. Aluminium and Fe (plus insoluble trace elements, such as Cr) become enriched in this zone. The formation of soil continues the process of enrichment in clay minerals and Fe oxides, although the formation of near surface calcrete/sulfates in the soil may dilute other components.

Quartz and muscovite are retained through the weathering cycle. Figure 6.6 depicts a generalised weathering profile on mafic rocks from Mt Magnet (WA), showing the mineralogical variation.

Weathering of Cenozoic basalts (containing olivine, pyroxene and glass) under more temperate conditions in south-eastern Australia, has produced

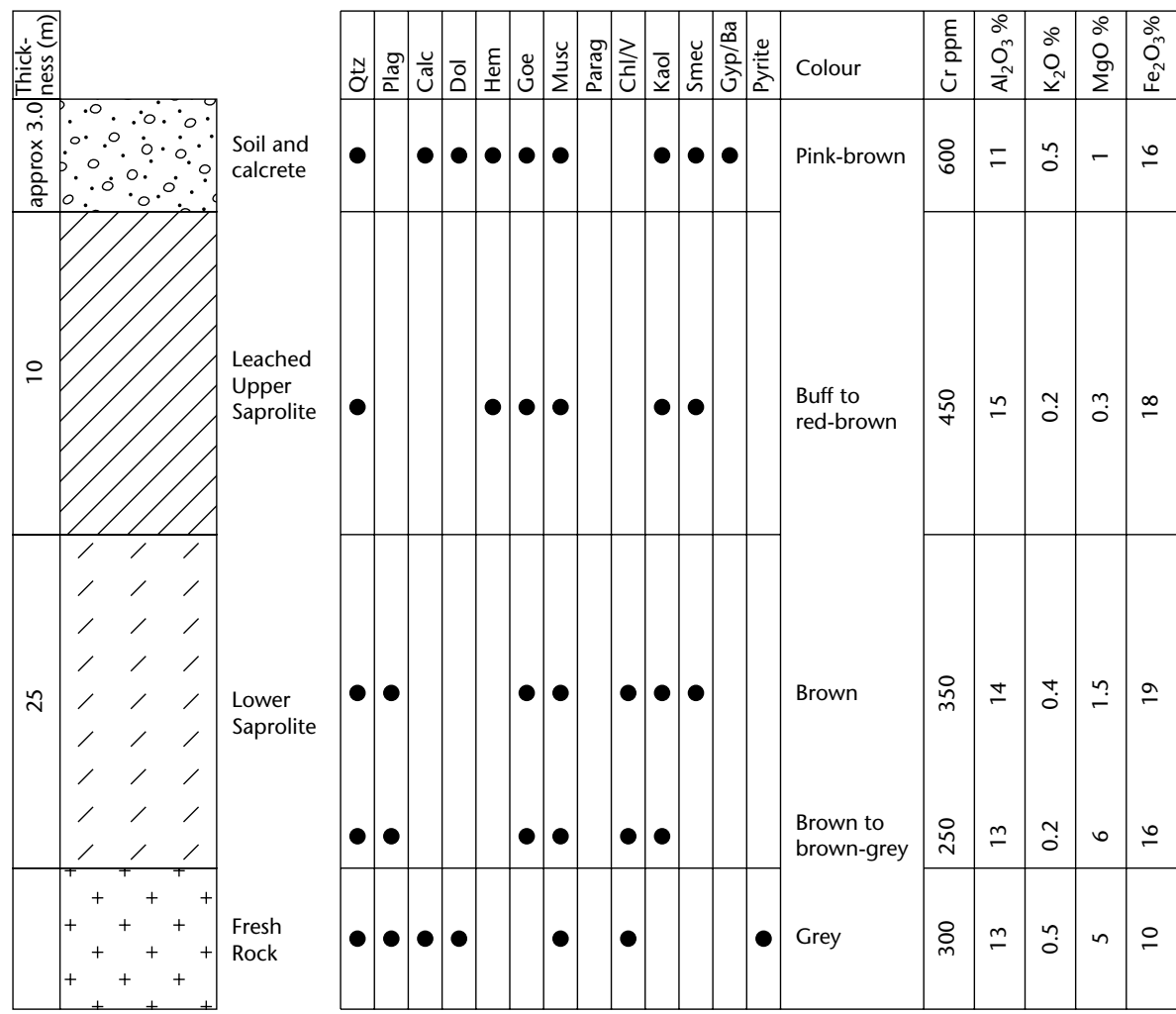


Figure 6.6: A generalised profile through weathered mafic rocks, Mt Magnet, WA (after Scott and Martinez 1990).

similar suites of minerals and elemental depletions/enrichments in saprolite and soil (Loughnan 1969; Moore 1996). The importance of water in weathering is illustrated by comparison with the weathering of basalts on the Moon where only physical, rather than chemical, weathering occurs (Section 14.2).

6.3.3 Weathering in ultramafic rocks

Ultramafic rocks consist mainly of olivine and pyroxenes, but they are commonly serpentinised or altered by metamorphism to amphibole–chlorite or talc-bearing assemblages. They are rich in Mg and poor in Al relative to other rock types. Chemical weathering of olivine- and serpentine-rich rocks produces smectitic clays, such as nontronite and saponite,

in the saprolite. These are altered to kaolinite in the upper saprolite, with loss of Mg. Some of this Mg may be precipitated as magnesite nodules in the lower saprolite. Iron – originally present in olivine and disseminated magnetite – forms goethite in the saprolite, which is dehydrated to hematite near the top of the weathering profile. Excess Si released during the weathering commonly forms deposits of opaline silica within the profile. Under free-draining conditions in high-relief terrains, Ni released from the weathering of olivine and serpentine can substitute for Mg in serpentine close to the weathering front to produce garnierite. In low-relief and poor-draining environments Ni may be taken up in smectitic clays or goethite in the lower saprolite.

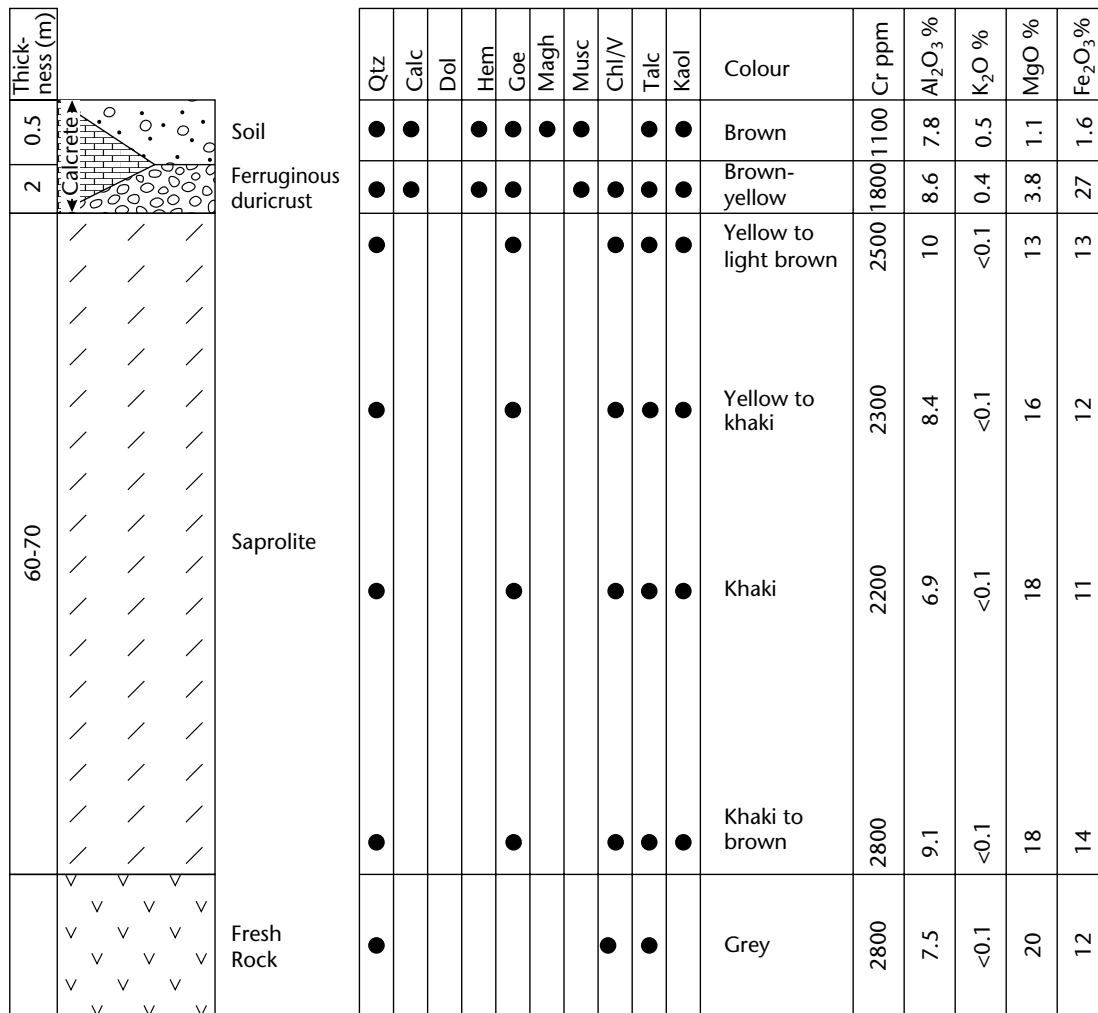


Figure 6.7: A generalised profile through weathered ultramafic rocks, Panglo, WA (after Scott 1990).

Significant concentration of Ni by these mechanisms can produce lateritic nickel deposits.

Unweathered serpentinised ultramafic rocks at Panglo (Yilgarn Craton of WA) are dominated by chlorite and talc, with only trace quartz (Figure 6.7). With the commencement of weathering, chlorite readily breaks down to form goethite, kaolinite, vermiculite and smectite in the lower saprolite. Thus, although the Mg in talc is retained, Mg from chlorite is gradually depleted up the profile. Aluminium and Fe from chlorite are retained in the neo-formed clays and Fe oxides. Ferruginous duricrust above the upper saprolite is particularly Fe-rich, indurated by calcrete, and commonly incorporates a transported component (indicated by the presence muscovite) so that Mg

and Cr contents are diluted. The soil formed above ferruginous duricrust shows further dilution of the diagnostic ultramafic components (Mg and Cr), although these are still present in significant amounts. Chromium present in spinels is retained during weathering, but that in chlorite (up to 3500 ppm) is transferred to goethite (8700 ppm) formed from the chlorite breakdown (Scott 1990).

6.3.4 Weathering in detrital sedimentary rocks

Detrital sedimentary rocks, and their low-grade metamorphosed equivalents, contain mainly quartz, muscovite, chlorite and, in some cases, plagioclase, carbonates, epidote and minor sulfides, together

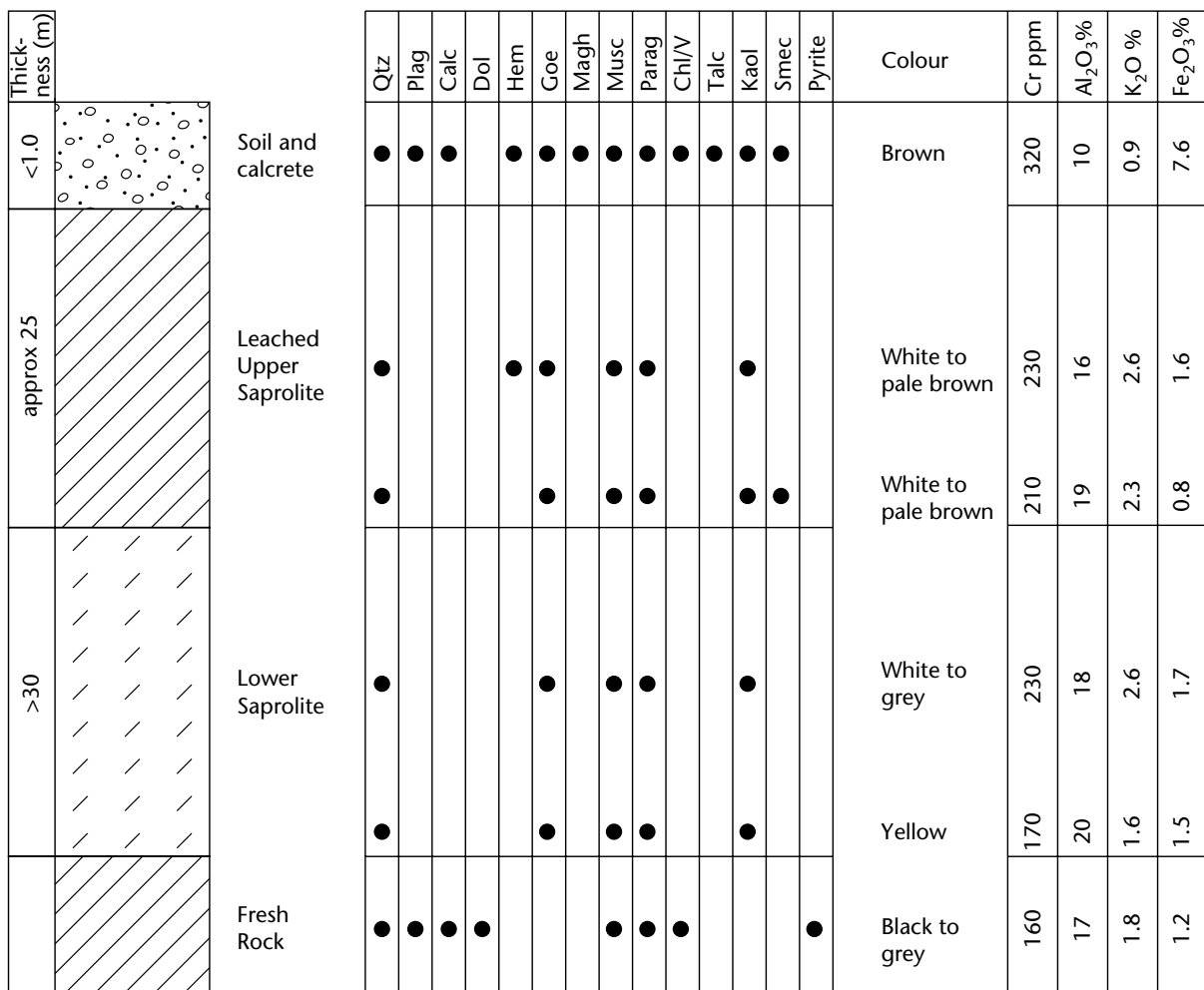


Figure 6.8: A generalised profile through weathered shale, Panglo, WA (after Scott and Dotter 1990).

with trace amounts of accessory minerals (zircon, rutile, ilmenite and apatite) and carbonaceous matter. As many of the accessory minerals are the weathering products of other rocks, they are less affected by further chemical weathering under a new weathering regime. In profiles on weathered sedimentary rocks (especially sandstones), it can be difficult to distinguish the weathered and unweathered zones. Physical processes generally play a more important role in the development of weathering profiles on these rock types – including increased fracturing and destruction of the rock fabric by stress relief, drying and wetting, freeze/thawing and veining by secondary minerals (such as Fe oxides).

Mottling and strong bleaching are other clues to identifying the weathered zone.

In unweathered Archean shale (interbedded with mafic volcanic rocks) in the Panglo area (Yilgarn Craton of WA), the major minerals are quartz, plagioclase, mica (muscovite + paragonite), carbonates, chlorite with or without pyrite. Weathering commences along cleavage/bedding planes in the shale, with the carbonates, pyrite and feldspar breaking down to form goethite and kaolinite. After this initial mineralogical change there is little further change throughout the lower saprolite (Figure 6.8). In the upper leached saprolite, smectite and hematite are developed, but chemical compositions are not mark-

edly different from those in the lower saprolite, except for residual concentration of Cr in neo-formed goethite and resistate rutile (Scott and Dotter 1990). The soil – with higher Fe but lower K and Al than in the leached upper saprolite – probably has a transported component at Panglo (Figure 6.8).

Non-Archean shales tend to have higher K contents than their Archean equivalents. The K may be present as K-feldspar and/or biotite, both of which weather out leaving lower K through the saprolite (for example, Dickson and Scott 1997). Chromium contents in shales in less mafic environments may be lower than at Panglo. Commonly, up to 1% organic matter may be present in black carbonaceous post-Proterozoic shales, with higher abundances in mineralised shales (Appendix 2). This organic matter (kerogen) has variable H/C and O/C ratios (Saxby 1976) which affects its weathering susceptibility, but ultimately it is broken down during weathering (Section 7.3.1) to produce a pallid saprolite: that is, black shales weather to white saprolite.

In arenaceous rocks, the feldspars and phyllosilicates (micas, chlorite) weather to kaolinite and goethite/hematite, but the abundant quartz component and accessory resistates remain largely unaffected. Quartz-rich sandstones break down mechanically, but do not change chemically during weathering – apart from the formation of minor Fe oxide coatings in fractures/voids and, in some cases of extreme weathering, precipitation of additional secondary silica. There are often silty interbeds, which weather as described above.

6.3.5 Weathering in carbonate-rich rocks

Limestones contain abundant calcite, with only minor silica/quartz and clays. Weathering involves dissolution of the calcite – with the loss of the Ca – and accumulation of the insoluble components to form a thin soil. (Dolomite tends to dissolve less rapidly than calcite, but, when it does, it frees Ca and Mg.) If there are significant clay and silt impurities or interbeds in the carbonate rock, these can be weathered and residually concentrated to form thicker (up to 30 m) accumulations of smectite, kaolinite or bauxite (gibbsite/boehmite). Minor Fe and Mn released from the carbonates during dissolution are generally oxidised to

form insoluble goethite, hematite and Mn oxides. The Fe oxides produce deep red-coloured residual soils known as *terra rosa*.

Profiles on carbonate-rich rocks are very distinctive: with sharp contacts between the unweathered (undissolved) rock and the weathering residue (Robertson *et al.* 2006). Unless there is a significant, non-carbonate component, there is generally no saprolite. Percolation of water along fractures and accompanying dissolution of the carbonate may lead to spectacular karst topography, with jagged and pinnacled rock surfaces forming the weathering front (Figure 6.9). Quartz, residual clays, insoluble Fe and Mn oxides and wind-blown dust commonly accumulate in hollows and cavities in the unweathered limestone and may form irregular and thick deposits. Zirconium, Ti and Si – present as minor components in the original rock – are residually concentrated, although much of the quartz and clay is derived from the more silty interbeds. Other primary constituents may be residually concentrated (such as, at Laowanchang in Guizhou Province, China, gold has been concentrated in this way). Further concentration of resistate minerals/elements occurs during the formation of the soil. The presence of chlorite in the soil (Figure 6.9) also makes the presence of an introduced component in the soil at Laowanchang obvious.

6.4 EFFECTS OF EROSION AND DEPOSITION ON PROFILE DEVELOPMENT

The weathering profiles described above are mature profiles for which there has been sufficient time and/or intensity of weathering for the zones to become well developed. However, in the case of recent exposure, the profiles may be immature, with only some features of the saprolite present and little or no surficial ferruginisation and soil development.

Mature weathering profiles may be partly eroded and a new weathering profile developed across the unconformity. In inset valleys (paleovalleys), clay-rich sediments commonly develop ferruginous megamottles (more than 100 mm across) as a result of later weathering (Ollier *et al.* 1988; Figure 6.10). These megamottles are formed by the mobilisation and concentration of Fe oxides often around roots or

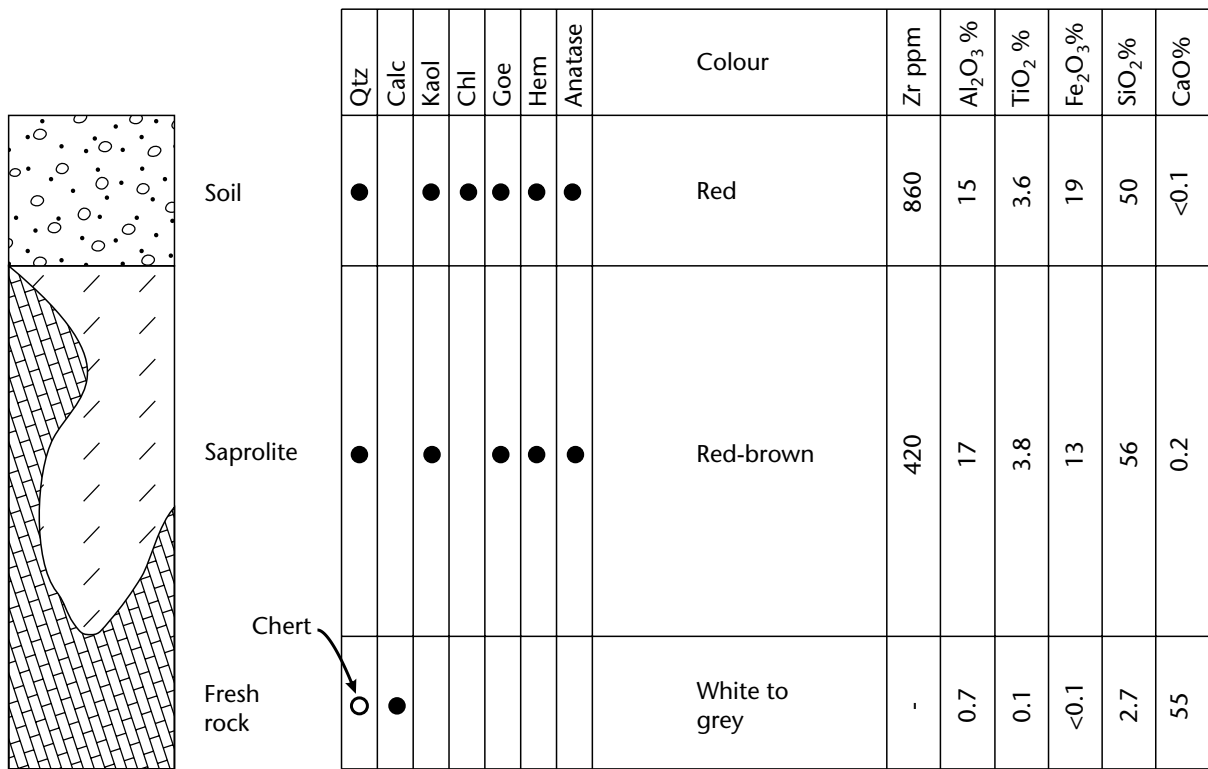


Figure 6.9: A generalised profile through weathering limestone, Laowanchang Au deposit area, Guizhou Province, China (data from J Mao *pers. comm.* 2007)

around shrinkage cracks (Anand 2001; Johnson and McQueen 2001).

Commonly, after a period of erosion – especially where a channel has developed – renewed deposition may commence, with accumulation of vegetable matter, which, in some cases, is pure and thick enough to ultimately become lignite, which is enriched in elements such as Mo and V. Such a reducing horizon can cause precipitation of secondary sulfides and U mineralisation such as at Mulga Rock, WA (Figure 6.11; Douglas *et al.* 2005).

The addition of transported material to weathering profiles can particularly affect the chemical composition of the soil layer. Most soils are affected by some colluvial movement of adjacent material (metre scale movement), but, unless this movement introduces material derived from a different rock type or from material with a different degree of weathering, it is unlikely to be recognised. Aeolian additions to profiles are obvious where they are present as sand

dunes, but they may not be so obvious when the introduced material is finer grained silt and clay. This extraneous component can also be worked into the profile by soil churning or bioturbation (Sections 3.4.4 and 8.3.2). This material can be recognised in the soil by detailed analysis of the particle size distribution, particle morphology and chemical and mineralogical characteristics (Scott 2005; Dickson and Scott 1998; Tate *et al.* 2007).

6.5 WEATHERING REGIMES

Chemical and physical conditions in the regolith can change significantly through time particularly in response to erosional, climatic, hydrologic and biological changes. At any one location, this can be the result of crustal plate movement, global or local climate change, as well as local tectonic and geomorphic factors. Regolith that has formed and been preserved over a long time span may therefore have

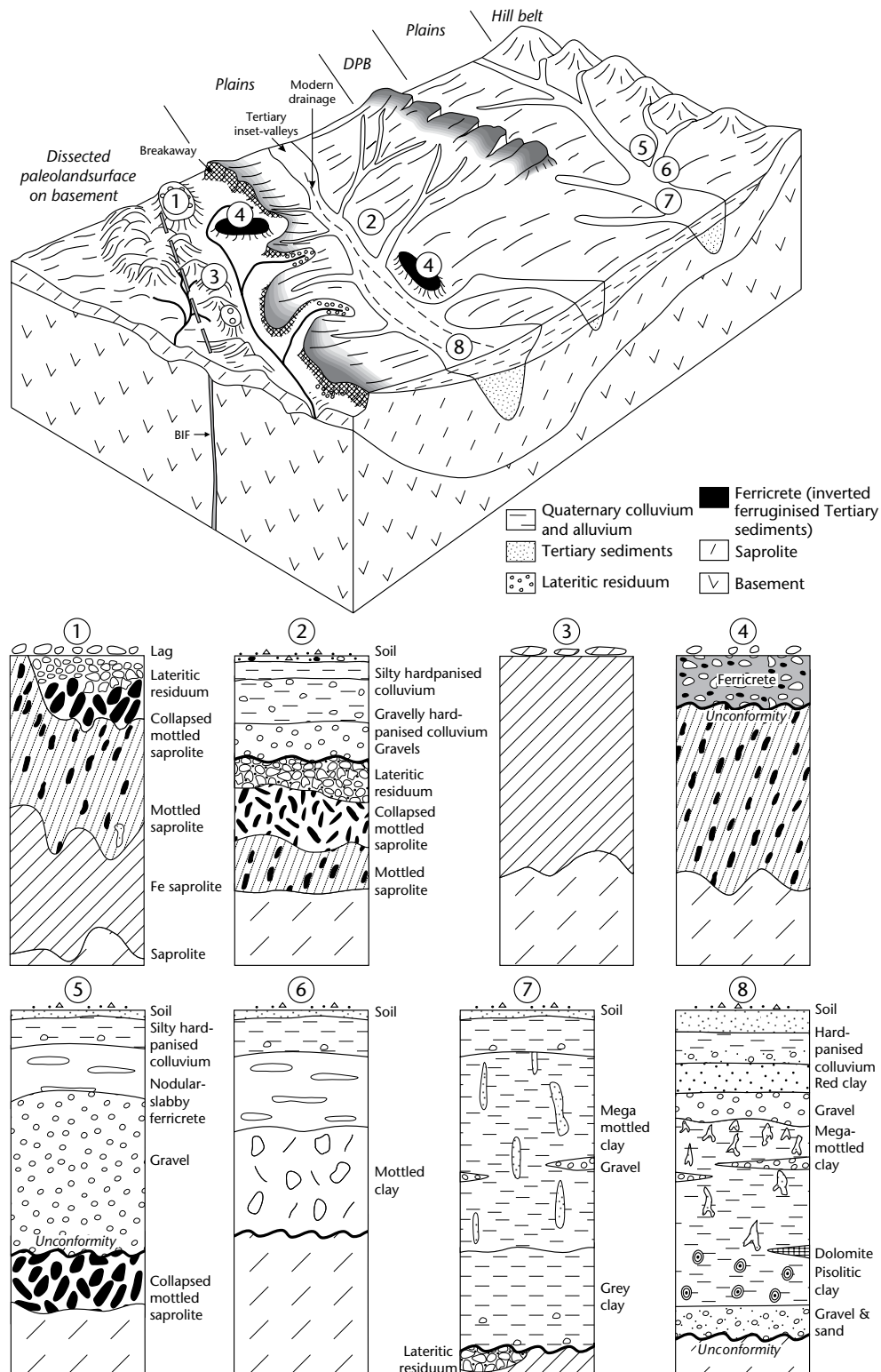


Figure 6.10: Formation of mega-mottles as the result of multiple weathering events (after Anand 2001),

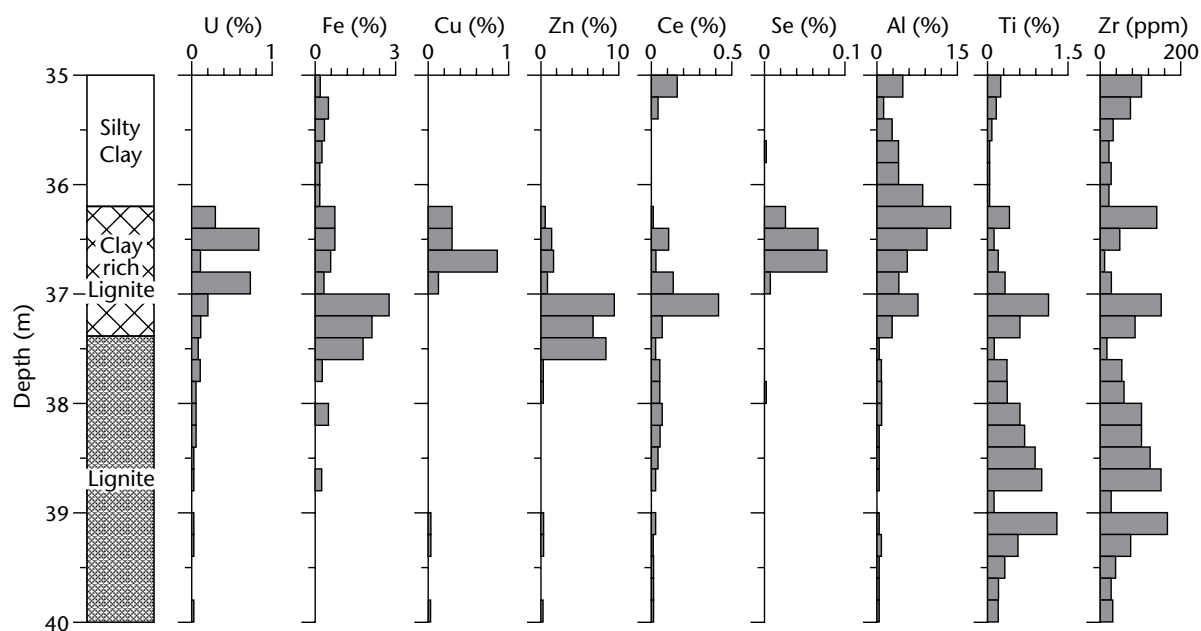


Figure 6.11: Development of a reducing environment leading to sulfide and U precipitation in clay-rich lignite, Mulga Rock, WA (after Douglas *et al.* 2005).

experienced a number of different weathering regimes. It is important to remember that not only is weathering a continuous process, but also the rate and type of weathering can vary depending on the stability of the minerals in the regolith and underlying bedrock under the conditions of the prevailing weathering regime. Changes in weathering conditions can have important implications for the dispersion of target and pathfinder elements used in geochemical exploration. Significant changes result in complex, multi-stage geochemical dispersion effects that may not be fully explained by the current weathering conditions. Patterns from different weathering regimes may be preserved in different parts of the landscape or they may be superimposed – depending on the history of landscape development, relative uplift, erosion and deposition. Figure 6.12 shows how the superimposition of a second weathering regime may largely obscure the effects of initial weathering.

6.5.1 Effects of long-term climate change

Climate is a major control on weathering regimes and, for most weathering profiles, climatic conditions have changed significantly during their formation (see

Chapter 2). The simplest case is for two markedly different climates, but, for very old weathering profiles – particularly in some Proterozoic and Archean terrains – the climatic conditions and weathering processes have been much more complex. Unravelling this complexity requires detailed studies using techniques such as paleomagnetic or isotopic dating (McQueen *et al.* 2007; Anand *et al.* 1997). For example, over much of southern Australia, major climatic variations through the Late Cretaceous and Cenozoic generally resulted in early deep chemical weathering under predominantly warm humid conditions (with high availability of organic material) onto which has been superimposed drier chemical weathering under increasingly arid conditions since the Late Cenozoic. The detailed picture is more complex, with fluctuations to at least two cooler-dry episodes before the Oligocene (McGowran and Li 1998). Paleomagnetic dating of ferruginous mottling in weathering profiles across the region indicates a number of periods of major hematite fixation – reflecting episodes of major oxidation and profile drying following intensive chemical weathering (Pillans 2006; McQueen *et al.* 2007; Smith *et al.* in press). These dated periods occurred during

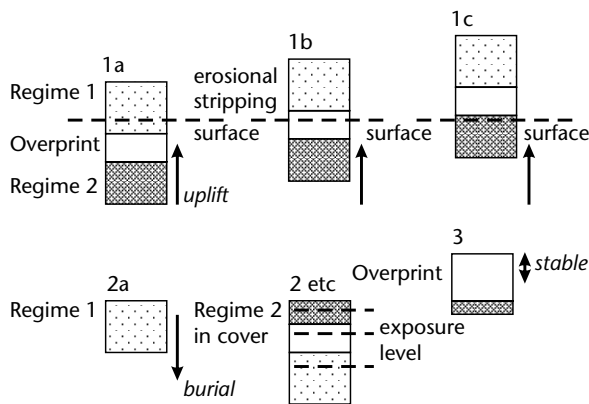


Figure 6.12: Superimposition of different weathering regimes.

the Paleocene (*circa* 60 Ma), Miocene (*circa* 16 and 12 Ma) and at some sites in the Late Cenozoic (*circa* 5 Ma). In some profiles, there is also evidence of earlier deep oxidation episodes, including in the Jurassic and Carboniferous (Pillans 2006). See Chapter 2 for further discussion of dating regolith materials.

6.5.2 Geochemical dispersion under different weathering regimes

Where weathering has occurred under a wide range of contrasting conditions, some minerals formed under an early weathering regime have a stability range that covers the conditions of later weathering. Thus kaolinite, Fe oxides, quartz and secondary silica will persist over a very wide range of weathering conditions and may retain minor and trace elements that are incorporated into their structures or concentrated in occluded accessory and secondary minerals. However, weakly adsorbed elements can be further dispersed under changed chemical conditions (particularly changed water activity, Eh and pH). Less-stable weathering products – such as smectitic clays, carbonates, sulfates, halides that are stable under one weathering regime – may be completely broken down under different conditions, releasing their contained major, minor and trace elements.

The availability of water under different weathering regimes is the major control on the resulting regolith geochemistry, element dispersion and fixation. Weathering regimes can therefore be considered as part of a spectrum from high water availability (wet,

mainly tropical and temperate, typically with annual rainfall greater than 500 mm) to low water availability (dry, mainly arid to semi-arid with annual rainfall less than 500 mm, or frozen). Seasonal fluctuations from wet to dry are also a common type of weathering regime in the subtropics.

Wet conditions – characterised by high water tables and a typically abundant and active biota – tend to favour hydration/hydrolysis reactions and mobility of reduced species, such as Fe^{2+} , and Mn^{2+} . Groundwater pH conditions are generally neutral to acid – the latter particularly in zones with abundant decaying organic material and around oxidising sulfides. High organic content promotes organo-complexing of many elements (Section 7.5). Most major elements are strongly leached – resulting in relative enrichment in Al, Fe and Ti. Trace elements are also strongly leached – particularly Cu, Zn, Cd and Ag – but some, including As, Bi, Cr, Mo, Pb and V, are retained or relatively concentrated in Fe-rich zones where these develop (Lecomte and Zeegers 1992).

Dry conditions typically favour oxidation reactions and a change to more complex, groundwater compositions, particularly with higher salinity, increased activity of carbonate and sulfate and regional neutral to alkaline pH. Some major elements (such as Ca, Mg and Na) may be retained in smectitic clays or enriched by evaporative concentration and re-precipitation in the regolith. More alkaline conditions favour mobility of trace elements such as As, Mo and U. Chloride and thiosulfate complexing can increase the solubility and mobility of some elements, such as Au, whereas other elements – especially Pb, Ag, Ba and Hg – become relatively fixed as insoluble chlorides and sulfates. Marked pH gradients around sulfide deposits that continue to weather generally results in dispersion of elements such as Cu and Zn to form distinct secondary haloes.

Weathering conditions in the Yilgarn Craton of Western Australia have spanned the range from warm humid conditions in the Late Mesozoic and Early Cenozoic to arid conditions in the Late Cenozoic (Chapter 2). There early-formed lateritic profiles – which are characterised by a modest amount of strong element leaching in the upper saprolite – have been modified under the more arid conditions, with the leached saprolite becoming more extensive as the water

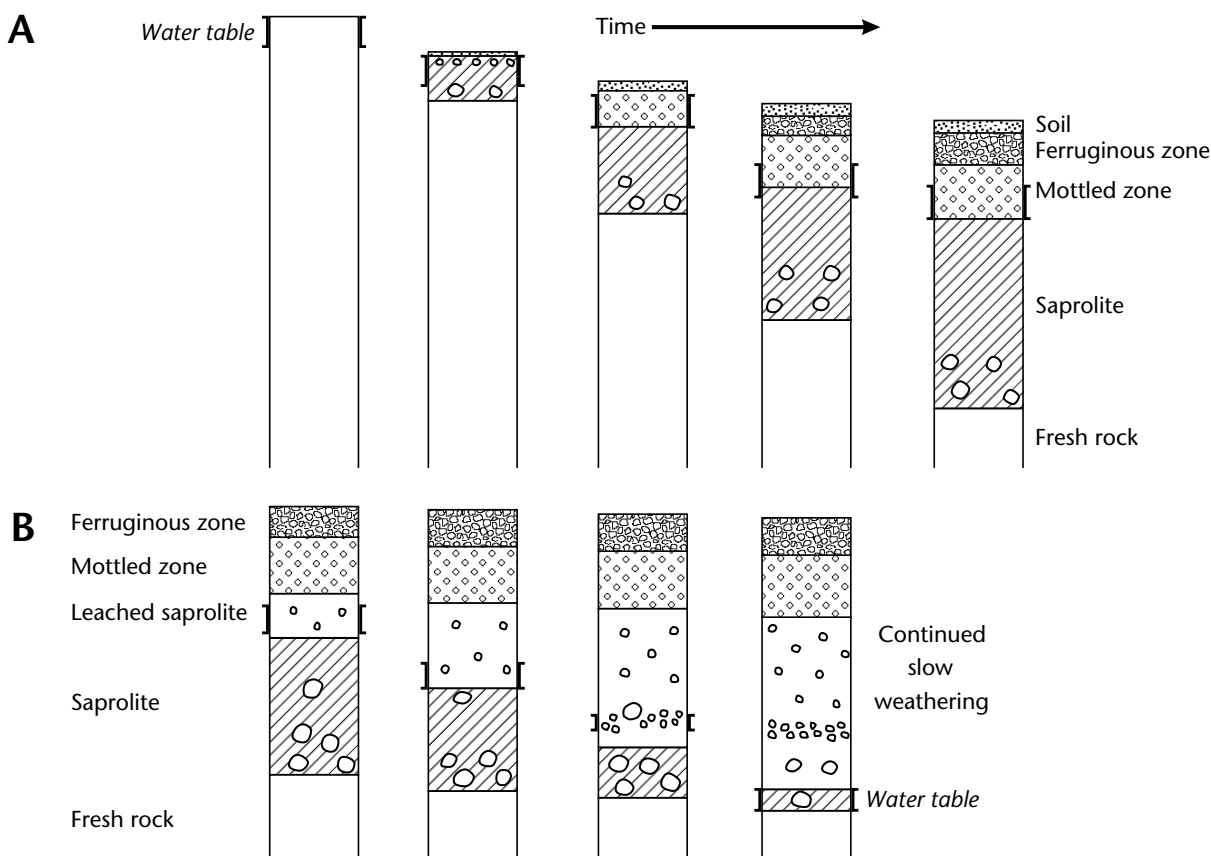


Figure 6.13: Effect of aridity on lateritic profiles of Yilgarn Craton (after Butt 1989).

table fell (see Figure 6.13). Under these conditions, the groundwaters became quite saline (Gray 2001).

6.6 DISCRIMINATING PARENT ROCK TYPES

In deeply weathered terrains, it is generally difficult to establish the parent rock type for regolith materials because of the profound mineralogical and chemical changes related to weathering and, in some cases, to secondary cementation and induration. In saprock and saprolite, the original rock fabric may be preserved to give clues to the original parent. In many cases, however, these features are destroyed – particularly in the upper part of the weathering profile. A number of studies have investigated the use of geochemical criteria – based on the least mobile elements – to discriminate parent rock types for *in situ* regolith.

The most widely used technique employs Ti/Zr ratios to discriminate between regolith derived from igneous rock types over the range from mafic to felsic (Figure 6.14; Hallberg 1984). These two trace elements are typically concentrated in rutile and zircon – minerals that are highly resistant to weathering – and are thus considered to be relatively immobile. (Some Ti may also be present in ilmenite, sphene, micas, amphiboles and pyroxenes and, as these minerals weather, Ti is released, but re-precipitated as stable anatase [TiO₂], which is effectively immobile (Butt 1985)). Furthermore, because Hf is more concentrated in the zircon of more fractionated igneous rocks, the Zr/Hf ratio in zircon can also be used to gain information about the parentage of highly weathered material (Figure 6.15). Minor and trace element contents of resistant minerals are also useful as guides to mineralisation (for example, Scott and Radford 2007).

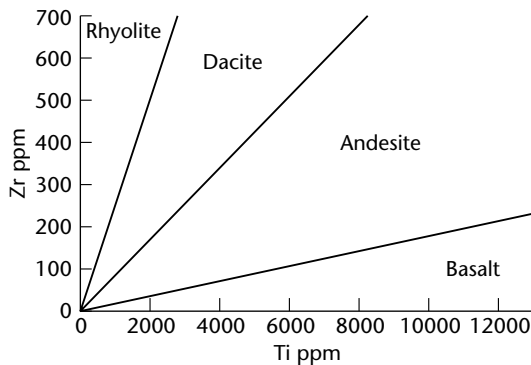


Figure 6.14: Ti/Zr plot for discriminating weathered volcanic rocks (after Hallberg 1984).

Of the major elements, Al is generally the least mobile and plots of Al_2O_3 - TiO_2 -Zr can also be used to discriminate regolith from different parent rocks. Including Al allows characterisation of regolith for a wider range of rock types, including sedimentary and metamorphic rocks (for example, Garcia *et al.* 1994; Figure 6.16). Interpretation of compositional variation for these three elements is based on the premise that sedimentation involves weathering, transport, mixing from different sources and sorting. In the first three processes, the contents of the less-soluble elements, such as Al, Ti and Zr, may vary in response to the degree of leaching of the soluble elements. However, their relative proportions are transferred from the source area into the bulk sediment or regolith without, or with little, modification. This material is then sorted according to the hydraulic properties

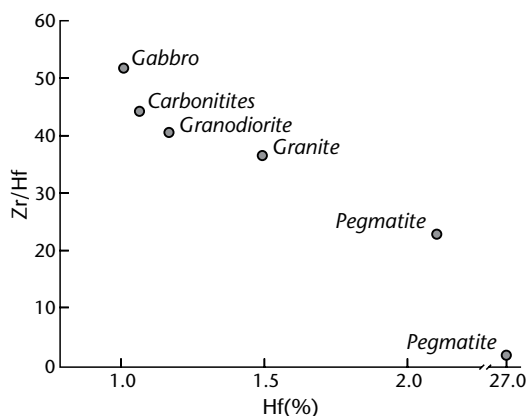


Figure 6.15: Zr/Hf for zircons in igneous rocks (after Černý *et al.* 1985; Wang *et al.* 1996).

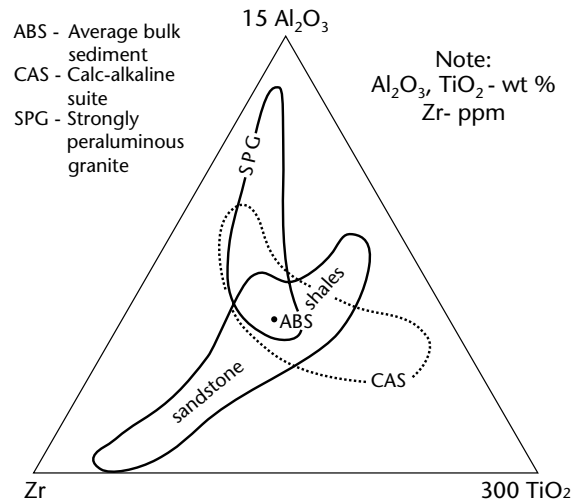


Figure 6.16: The $15\text{Al}_2\text{O}_3$ - 300TiO_2 -Zr weathered rock type discriminant diagram (after Garcia *et al.* 1994).

of its mineral components to produce a chemical fractionation between complementary shales and sandstone (Figure 6.16). Other rock types, such as felsic and mafic igneous rocks or immature volcanic-derived sediments, will also plot in specific fields.

6.7 CHARACTERISING AND IDENTIFYING REGOLITH MATERIALS

Major chemical changes that occur during weathering – particularly progressive loss of Na^+ , K^+ , Ca^{2+} , Mg^{2+} (some Si^{4+}) and retention of Si^{4+} , Al^{3+} and Fe^{3+} (Section 5.4.2) can be used to help to characterise regolith materials that are at different stages of mineralogical/chemical evolution. The K/Al and Mg/Al ratios can commonly be used in this way by comparing two mobilised elements with a relatively immobile major element. Figure 6.17 shows an example of the compositional distribution in terms of this index for three different types of regolith found in both the Kalgoorlie region of Western Australia and the Cobar region of eastern Australia (Johnson and McQueen 2001; McQueen 2006). The three different regolith materials in these examples are *in situ* saprolite/saprock, lacustrine clays (difficult to distinguish visually from weathered saprolite) and younger ferruginous alluvium/colluvium. These materials have different regolith histories that are reflected in their

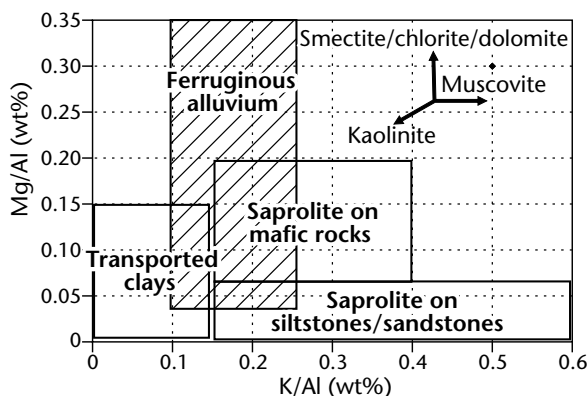


Figure 6.17: Compositional variability of regolith in terms of Mg/Al versus K/Al (after McQueen 2006).

mineral and chemical composition. The lacustrine clays were deposited in the Late Mesozoic to Early Cenozoic and appear to have been derived from a deeply weathered landscape. They were well sorted during erosion and transport to produce a kaolinite–smectite–quartz dominant sediment. The saprolite/saprock contains significant muscovite and illite, which have been variably altered to kaolinite – depending on degree of *in situ* weathering and depth in the profile. The younger alluvium/colluvium was deposited in the Late Cenozoic when climatic conditions were significantly drier and chemical weathering less intense. Erosion of less altered profiles, more limited sorting and low levels of post-depositional weathering produced material that retained significant amounts of weakly altered phyllosilicates. These sediments can also contain local concentrations of dolomitic calcrete, which will be apparent in their Mg/Al ratio. There is some compositional overlap between the three materials, and the clearest distinction is between saprolite and transported clays. This approach to chemically distinguishing regolith types can be applied within particular regions where there are regolith components characterised by different parent rock compositions, different degrees of weathering or different histories of sorting or remixing/homogenisation during transport. The differences can be established with an orientation survey.

Trace element characteristics – particularly for those elements hosted by resistate minerals – can be useful for identifying different types of regolith. For

example, regolith derived from mafic or ultramafic rocks typically has higher contents of Ti, V and Cr – hosted in rutile/anatase, ilmenite, magnetite and chromite – than regolith derived from felsic or meta-sedimentary rocks. Aeolian material in the regolith may have high Zr (and Hf) contents relative to *in situ* material due to the greater abundance of zircon grains in the introduced material (Tate *et al.* 2007).

6.8 DEGREE OF WEATHERING AND WEATHERING HISTORY

The Chemical Index of Alteration (CIA) has been widely used to quantify degree of rock weathering (Nesbitt and Young 1982). This index, $CIA = 100Al_2O_3 / (Al_2O_3 + CaO + Na_2O + K_2O)$, reflects the breakdown of feldspars and mica to kaolinite, but it has a major drawback in that it estimates the total history of chemical weathering from the primary source rock: that is, including that already present in sedimentary rocks subjected to further weathering. It is thus difficult to apply this index as a direct measure of the *in situ* weathering of a particular regolith sample. However, it can be useful for comparing samples within profiles developed on a variably weathered common rock type (Figure 6.18). Weathering trends within a profile, or

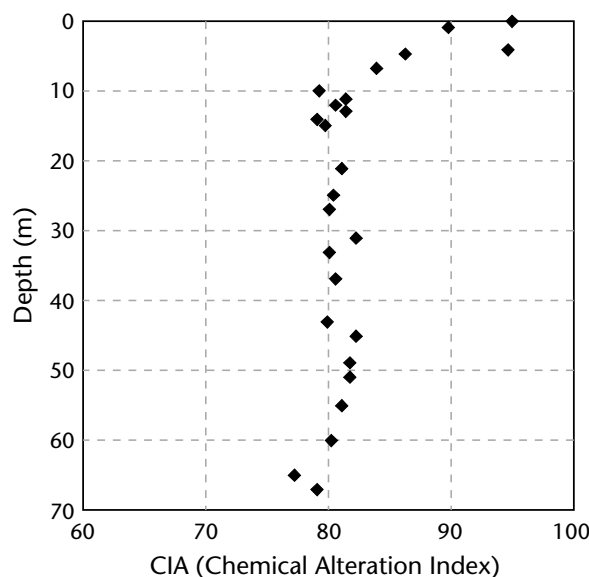


Figure 6.18: Chemical Index of Alteration (CIA) variation with depth in weathered metasediments, Cobar region (after McQueen 2006).

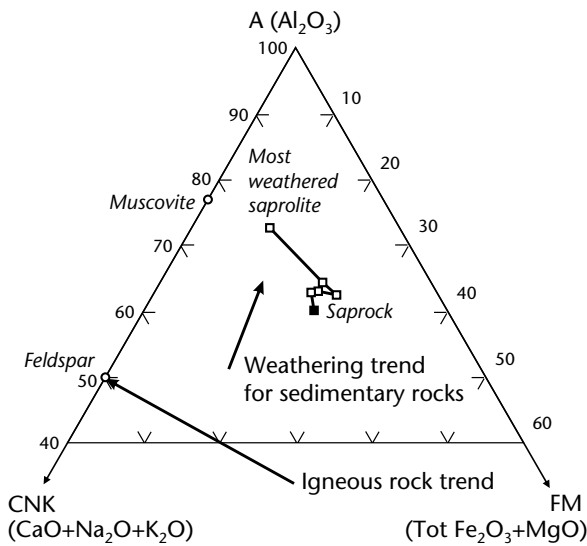


Figure 6.19: A-CNK-FM plot for a weathering profile through metasediments, Cobar region (after McQueen 2006).

group of profiles, on common rock types can also be distinguished using the relative mole proportions of Al_2O_3 , $(\text{CaO}+\text{Na}_2\text{O}+\text{K}_2\text{O})$ and $(\text{total } \text{Fe}_2\text{O}_3+\text{MgO})$ (Nesbitt and Young 1989). Thus an A-CNK-FM plot for samples from a single weathering profile within metasediments in the Cobar region (Figure 6.19) reflects the relative increase in Al and loss of Fe, Mg and alkalis as weathering proceeds. This approach is complicated by the inclusion of Fe, which can be accumulated at redox boundaries or rapidly mobilised from Fe-bearing carbonates in weathering sedimentary rocks. Formation of secondary dolomite can also complicate Mg contents of strongly weathered regolith.

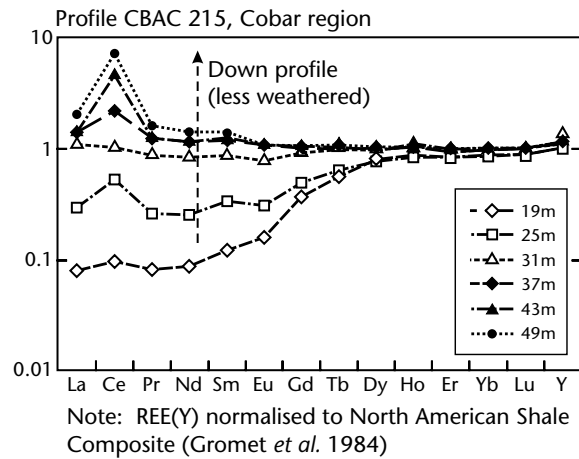


Figure 6.20: REE distribution in a weathering profile through metasediments, Cobar region (after McQueen 2006).

The REE are commonly used as petrogenetic indicators because of their similar chemical properties, typically low solubilities and assumed resistance to fractionation in supracrustal environments. However, a number of studies have shown that under some weathering conditions REE are significantly mobilised and fractionated (for example, Nesbitt 1979, Duddy 1980, Sharma and Rajamani 2000, Gray 2001, McQueen 2006). Far from being immobile, REE can be significantly redistributed during weathering to the point where their depletions in one part of the profile, and subsequent enrichments in other parts, can provide an index for the intensity and style of chemical weathering. Analysis of REE distributions through deeply weathered profiles in the Cobar region indicates significant leaching in the upper

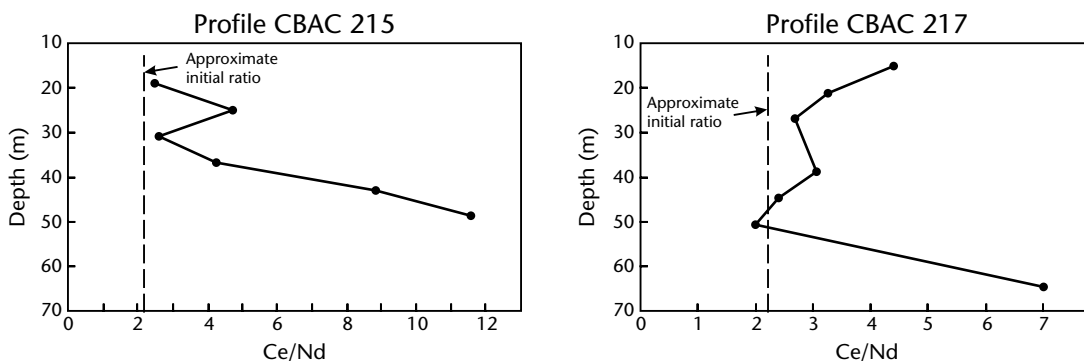


Figure 6.21: Ce/Nd plots for weathering profiles through metasediments, Cobar region (after McQueen 2006).

parts, and enrichment in the lower zone close to the weathering front (Figure 6.20). The light rare earth elements (LREE) – particularly Ce – appear to have been the most mobile under the weathering conditions that pertained in this region. This pattern is less marked in the least-weathered profiles and there is also some indication that profiles in different settings have different patterns of REE mobility and enrichment (McQueen 2006). The REE may therefore provide a basis for evaluating the extent of chemical weathering and leaching. Comparison of Ce (the most mobilised LREE) with Nd (another LREE with similar properties and initial distribution) is one possibility for a REE mobility and weathering index. Examples of Ce/Nd plots for different weathered profiles in the Cobar region are shown in Figure 6.21. Departure of the Ce/Nd ratio from the initial ratio (typically around 1.5 to 2.5 for sedimentary rocks) indicates the relative depletion or enrichment of Ce and provides an indication of chemical leaching of the rocks.

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