1A. ON THE ORIGINS, GEOMORPHOLOGY AND SOILS OF THE SANDPLAINS OF SOUTH-WESTERN AUSTRALIA

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INTRODUCTION

SANDPLAINS ARE PROMINENT ELEMENTS IN the surface geology and geomorphology of Western Australia, where they provide a distinctive substrate with characteristic soils and weathering profiles. They occur extensively throughout the northern and southern parts of Western Australia, appearing in the major sedimentary basins, the adjoining cratons, and along the coastal margins. In northern parts of Western Australia, sandplains have a striking expression in the extensive dune fields of the Carnarvon and Canning Basins. In south-western Australia their geomorphological expression is more localised and less dramatic, yet they remain prominent elements in the landscape (Beard & Sprenger, 1984) and support the distinctive kwongan vegetation (Pate & Beard, 1984; chapter 2).

The sandplains associated with kwongan vegetation can be grouped into inland sandplains and coastal heath sands (Bettenay, 1984). Early mapping by Prescott (1931) and Northcote (1960) identified the broad extent of the sandplains. The mapping of soils and landscapes in the rangelands began in 1953 by CSIRO (Speck *et al.*, 1960). In 1982 the Western Australian Department of Agriculture began a regional-scale mapping program in agricultural areas, which was completed in digital form in 2003 (Schoknecht *et al.*, 2004). The land system mapping and attribution DAFWA Soil Landscape Database (2007) provides a comprehensive description of the sandplain soils and their distribution (Newsome, 2000).

Bettenay (1984) provided an overview of the relevant historical literature of the sandplains with a pedological emphasis and readers interested in the early literature are referred to his review. Bettenay (1984) did not include the important work of Carroll (1939) and Prider (1966), but attention is drawn to this here (see below). Since the review of Bettenay (1984), more recent work relevant to a discussion of inland sandplains has appeared (*e.g.*, Glassford & Semeniuk, 1995; Ollier & Pain, 1996; Newsome, 2000) which can now be placed into a more comprehensive understanding of regolith development generally (Anand & Paine, 2002; Scott & Pain, 2008). Similarly, a more complete understanding of the coastal successions (Hewgill *et al.*, 1983; Murray-Wallace & Kimber, 1989; Kendrick *et al.*, 1991; Bastian, 1996; Price *et al.*, 2001; Hearty & O'Leary, 2008) allows a stratigraphy-based discussion of the coastal sandplains, and makes a chronostratigraphic approach to the associated soil development possible (Laliberté *et al.*, 2012). A further useful addition to an understanding of the pedology of the sandplains is the field catalogue of soils (McArthur, 1991), which provides 'type soils' sites for sandplain locations in south-western Australia.

This chapter discusses the origin and extent of the sandplains and their long-term stability in the context of the changing Late Cenozoic climate background. These are themes that have contributed to the Hopper's OCBIL theory (Old, Climatically-Buffered, Infertile Landscapes) (Hopper, 2009). Recent advances towards a more comprehensive understanding of weathering events and palaeoclimate, coupled with the development of a variety of dating techniques, have made a fuller, albeit still tentative, discussion of inland sandplains possible. Similarly, changes in sea level in the Tertiary and Quaternary are now better known (*e.g.*, Miller *et al.*, 2011), and, with the numerical age controls now available, provide a better guide to the origin and history of the coastal sandplains.

For a complete list of terms used in this chapter, the reader is referred to the Glossary for this chapter.

INLAND AND COASTAL SANDPLAINS

The geomorphological significance of sandplains

Inland sandplains are widespread in south-western Australia and are found in a range of landscape settings. In general terms, the inland sandplains reflect the long-term denudational history of the region, specifically the transport-limited erosion regime (*i.e.* the rate of supply of material exceeds the capacity of transport processes to remove it (Carson & Kirkby, 1972; Stallard, 1995). This regime has prevailed over the Cenozoic (and most likely longer time scales). In the Northern Hemisphere, ice-sheets removed much of the pre-Quaternary regolith cover. In contrast, south-western Australia has not experienced glacial events since the Early Permian, *ca.* 260 million years ago (Geological Survey of Western Australia, 1990). This absence of late Cenozoic glaciation, coupled to the tectonic stability of south-western Australia, has allowed the weathering product to be retained in the landscape.

These 'boundary' restraints provide the link with the OCBIL theory (Hopper, 2009). It is thought that in these landscapes natural selection has favoured limited dispersability of sedentary organisms, resulting in an elevated persistence of lineages and long-lived individuals, high numbers of localised rare endemics and strongly differentiated population systems (Hopper, 2009; p. 60). However, these claims need to be placed into the context of a lack of firm evidence as to the age of the sandplains and the climatic changes they have experienced since their formation. Inland sandplains, while essentially source-located (see below), have not simply acted as passive elements in the landscape. Some have clearly been reworked, albeit to differing degrees, during the Late Cenozoic. This has pedological implications that need to be incorporated in any model of the long-term development of plant–soil interactions. Also, large parts of the Northern Hemisphere were not glaciated during the Late Cenozoic and possess suitable source-lithologies, yet still lack the regolith characteristics of sandplains. This highlights the importance of the low-erosion potential that characterises the relatively low relief expression of much of south-western Australia, although certainly not as much as envisaged in some earlier studies (*e.g.*, Finkl & Fairbridge, 1979).

The occurrence of sandplains along the coastal margins of south-western Australia is the direct outcome of the global sea-level fluctuations of the last few million years. The sediment characteristics and response to weathering, coupled with relative tectonic stability, has allowed these sediment-associations to remain in their depositional settings. The coastal sandplains have a history that extends into the Pliocene, and there might be even older elements in these successions, perhaps extending to the Miocene (*i.e. ca.* 5.3 million years ago).

For much of the Perth Basin, the Quaternary part of the coastal succession consists of a series of well-defined coastal-barrier complexes associated with periods of high sea levels that extend into the middle Quaternary and provide an intriguing chronosequence of events, in which soil development, pedogenesis, and associated changes in nutrient status can, in theory, be traced through time.

The origin of inland sandplains

Bettenay (1984) termed the inland sandplains 'Lateritic Sandplains'. However, the term 'laterite' carries specific genetic and profile connotations (see discussion in Anand & Paine, 2002). So although inland sandplains are associated with laterites, the term 'lateritic sandplain' is avoided here in favour of the more generic term 'inland sandplain'. This also allows a distinction between sandplains in cratonic and sedimentary basins, emphasising the differences in the controlling geology.

Inland sandplains formed in cratonic and sedimentary basins share a common weathering regime and denudational history, both in their original formation, and also in the events that may have modified or reworked the deposits. However, given the more extensive sandplains over arenaceous lithology in, for instance, the northern part of the Perth Basin, the production rates of sand-dominant regolith are likely much greater on sedimentary lithologies (as noted by Beard & Sprenger, 1984).

The inland sandplains occur in a broad belt from Shark Bay to west of Esperance. The soils are typically yellow deep sands, pale deep sands (or Orthic Tenosols in the Australian system of soil classification), and yellow sandy earths (Orthic Tenosols, Yellow Kandosols). Soil depth can vary from a few decimetres to several metres. Profiles often contain ferricrete (a hard, erosion-resistant layer of soil cemented by iron oxides), or ferruginous or aluminous 'gravels'. In many instances, clay content increases with profile depth. The profiles are infertile (Bettenay, 1984; McArthur, 1991; Rogers, 1996) and have a limited ability to hold water (Bouwer, 1978; Scott *et al.*, 2012), both of which have important impacts on associated plant communities (Richards & Caldwell, 1987; Dawson & Pate, 1996; Lambers *et al.*, 2010; chapters 4 and 5). The inland sandplain soils contrast with the deep sandy soils of the forest and woodland areas in the region, and the sandy surface soils over shallow clay B horizons (duplex, or texture-contrast soils) of the mallee (*i.e.* small multi-stemmed eucalypt) communities.

Ultimately, the inland sandplains originate through the development of a regolith environment that provides the necessary weathering product. In a geological sense, a sandplain represents a first-cycle quartz arenite (*e.g.*, Johnsson *et al.*, 1988). This is clearly reflected in the quartz-dominated mineralogy, with most labile weathering mineral components absent (Carroll, 1939), which accounts for the low nutrient status of sandplain soils (chapter 4). The dominance of quartz is not surprising, given that a 1 mm quartz crystal at 25°C and associated with a weathering environment of pH 5, has a lifespan of 34 x 10^6 years (Lasaga *et al.*, 1994).

Carroll (1939) was able to show that sandplain source lithologies can be clearly identified. Granitic/gneissic bedrock and associated sandplains are characterised by an assemblage of zircon, rutile, garnet, sphene, tourmaline, and monazite. In contrast, meta-sedimentary lithologies and associated sandplains are typed by andalusite, staurolite, sillimanite, kyanite and spinel. The clear correspondence of mineral assemblage to underlying bedrock points to an essentially *in-situ* origin of sandplains. Much of the later work supports this inference (*e.g.*, Mulcahy, 1960; Prider, 1966; Brewer & Bettenay, 1973). Recent work by Newsome on the Victoria Plateau sandplain (Newsome & Ladd, 1999; Newsome, 2000; Newsome & Walden, 2000) reiterated these conclusions and has lent support to the general claims that the sandplains have an essentially *in-situ* origin with a degree of local reworking. This large body of work makes it difficult to accommodate the view of an extensive aeolian origin for inland sandplains (Glassford & Semeniuk, 1995). However, an essentially *in-situ* origin of inland sandplains does not preclude local reworking of sandplains into colluvial and alluvial registers, or even the occurrence of some large dunes attesting to significant mobility (see below).

Origins of coastal sandplains

The origin of coastal sandplains is linked with sea-level fluctuations over Neogene and Quaternary time scales. During the Miocene and Pliocene (*i.e.* between *ca.* 23 and 2.6 million years ago) there was a progressive lowering of sea-level punctuated by global high-stands which became subsequently dominated by the Milankovitch cycles during the Quaternary (Zachos *et al.*, 2001; Miller *et al.*, 2011). The details of sea-level change over the last *ca.* 150,000 years (Lambeck & Chappell, 2001; Yokoyama & Esat, 2011), including the Holocene (*i.e.* the last 11,700 years) (Lewis *et al.*, 2013), are now well established. These provide key stratigraphic anchors for understanding the chronology of coastal succession in south-western Australia.

Along the coast of south-western Australia, global sea-level changes are most clearly expressed in the Holocene succession, the Tamala Limestone and the Bassendean Sands, and with reference to the older Yoganup Formation (Kendrick *et al.*, 1991; Bastian, 1996). The Holocene succession is displayed strikingly in a number of beach–ridge/dune associations. The barrier structure of the Tamala Limestone is especially well displayed in the central Perth Basin. The Bassendean Sands are geomorphologically more weakly defined, but are recognised as representing a siliciclastic coastal dune succession, associated with the underlying marine facies of the Ascot Formation. It is noteworthy, that Noongar people showed a clear awareness of the differences between the Tamala Limestone terrains – Booyeembara – and the Bassendean Sands – Gandoo (Lyon, 1833).

In the Perth Basin, where the succession is best known, the detailed (local) soil associations have been typed in association with the Quindalup, Spearwood and Bassendean dune systems (see McArthur, 1991). In this classification the Quindalup dunes correspond to the Holocene succession and the Spearwood dunes are equivalent to the Tamala Limestone.

Associated with the Spearwood dunes/Tamala Limestone are extensive siliceous 'yellow sands'. There have been claims that these sands represent arid-zone aeolian sediment (Killigrew & Glassford, 1976), but this is inconsistent with other geological evidence (summarised in Tapsell *et al.*, 2003). Following this, and placed into the wider context of limestone weathering and karst formation, the yellow sands associated with the Tamala Limestone succession are interpreted as a residual weathering product, with local-scale reworking.

A related issue is the low carbonate status of the Bassendean Sands association and its significance both in terms of primary depositional environments and subsequent soil formation. In the subsurface, the Bassendean Sands are clearly associated with the Ascot Formation, and the extensive sand terrains associated with the Bassendean Sands are thought to represent a regressive dune facies. The sediments have little carbonate which is interpreted as due to low carbonate productivity or dilution by terrestrial siliciclastic inputs (Kendrick *et al.*, 1991).

AGE STRUCTURE AND MOBILITY OF SANDPLAINS

Deep weathering history and the long-term origin of inland sandplains

Generally, a long weathering history is advocated for the formation of the sandplains, but often without any firm age control. Given the difficulty of dating regolith, the absence of a convincing age framework for regolith sequences is not surprising. Some age control is offered by general geological considerations; for example, the extent of the Eocene transgression provides a maximum age for sandplains in the wider Stirling Range region. Pillans (2008) provides a comprehensive review of attempts to ascertain the ages of Australian regoliths. For shorter time scales, the development and application of optically stimulated luminescence (OSL) has made it possible to obtain reasonably reliable dates from sediment and soil successions that could not previously be dated (see Rhodes, 2011). In principle, OSL dates can extend back hundreds of thousands of years. The challenge that remains is to date regolith events that might extend into the Neogene and even older.

The region between Mingenew and Mullewa provides an example of the antiquity and stability of the inland sandplains. These sandplains are located north of the Dandaragan sandplain (Churchward, 1970) and are part of the wider Victoria Plateau sandplain (Newsome, 2000). The region is associated with a series of Permian sediments and has a geomorphological expression in which isolated mesas capped by thick sand (above a largely ferricrete surface associated with a deeply weathered profile) are isolated from the plateau margin (Fig. 1). While present rates of retreat of the scarp by erosion are likely high, a very considerable time would be required to attain such a geomorphological expression. A Miocene–Pliocene age for the onset of the sandplain formation seems therefore likely.

Figure 1. Geomorphological expression of elevated, remnant sandplain on a mesa of Permian sediments – Irwin Valley, north of Mingenew.

A more secure estimate for the formation of the sandplain on the wider Victoria Plateau is possible. On the basis of palaeomagnetic work, Schmidt & Embleton (1976) advocated a Late Oligocene to Early Miocene age (*i.e.* > 16 million years ago) for extensive 'laterisation' in this area, to which sandplain formation is clearly linked (Newsome, 2000). However, a revision of the Australian Apparent Polar Wandering Path is now thought to indicate a Late Miocene to Pliocene age, 6±4 Ma (Pillans, 2008), for such a deep weathering event. Pillans (2008) obtained a similar palaeomagnetic weathering age for mottled saprolite beneath bauxitic ferricrete at Jarrahdale, south of Perth. Other indications of a late Miocene age for deep weathering events is provided by radiometric dating (uranium–thorium/helium, or U-Th/He) of ferricrete nodules from the Darling Range (Pidgeon *et al.*, 2004). It seems likely that the necessary setting for sandplain development could well be linked to these Neogene events.

How stable are sandplains once formed? This answer is linked to estimates of long-term erosion rates. Much of the early discussion of this theme was highly speculative, with estimates of very limited erosion rates between 0.1 and 0.2 metres per million years (Fairbridge & Finkl, 1980). These claims were overturned by subsequent work, which advocated rates of 1.5–2 metres per million years during the Late Cretaceous and early Tertiary (Van de Graaff, 1981). More recent work using fission-track dating has pointed to low denudation rates of 1–2 metres per million years over the Late Tertiary (Kohn *et al.*, 2002). For the Yilgarn Craton, over longer-term Tertiary to Cretaceous time scales, rates more likely amount to 10–15 metres per million years (Kohn *et al.*, 2002). Such relatively high erosion rates would make it unlikely that sandplains have persisted over mid-early Tertiary to Cretaceous time scales. Given that semi-arid climates in Western Australia appear to date back to the late Pliocene (see below), the low erosion rates that would accompany drier conditions make it likely that the sandplains are a feature that can certainly extend back into the Pliocene and possibly the late Miocene. Older ages of the sandplains seem unlikely.

The timing of the onset of present climate regimes over south-western Australia is generally poorly constrained. Martin (2006) points to the possible onset of drier conditions during the mid-Miocene, while an early Pliocene sequence from Lake Tay in south-western Australia (Bint, 1981) recognised minor rainforest elements in association with Casuarinaceae and *Eucalyptus* sclerophyll woodland association. Dodson & Macphail (2004) recognised a number of arid events increasing in frequency at around 2.5 Ma in a Late Neogene lacustrine succession from the northern Perth Basin. Fujioka *et al.* (2005; 2009) provided further evidence for the onset of more arid conditions over central Australia between 2 and 4 million years ago (Fujioka *et al.*, 2005), with stronger arid conditions evident by about 1 Ma (Fujioka *et al.*, 2009). Chen & Barton (1991) provided an age of 0.9 Ma for the onset of full aridity in the Lake Eyre region. Tasman Sea dust deposition, indicating arid conditions over south-eastern Australia, is not evident before 0.45 Ma (Hesse *et al.*, 2004), which corresponds with lake-to-playa transitions in south-eastern Australia between 0.4 and 0.7 Ma (An *et al.*, 1986). For south-western Australia, Zheng *et al.* (1998) proposed the onset of full arid conditions in the Lake Lefroy region around 500,000 years ago; given the general view, this would appear to be too recent.

Given the requirements for sandplain formation, including weathering history, erosional regime (and rate), and general palaeoclimate context of the Late Cenozoic, we suggest that the sandplains of inland south-western Australia have a history that can be accommodated in a late Miocene to Pleistocene time scale. The sandplain terrains in their present expression – both in terms of geomorphology and arid climate – may be features that are no older than Pleistocene in age.

Numerical dating of sandplains and significance to sandplain stability over Quaternary time scales

Very little is known about the long-term stability of the Western Australian sandplains, or their responses to Late Cenozoic climate events. General geomorphological and sedimentological considerations demonstrate reworking, but when this occurred is rarely known. However, a cursory set of OSL dates from the wider Eneabba sandplain demonstrate that the upper 100 cm of the sandplain was remobilised between 19 ka and 31 ka (Krauss *et al.*, 2006).

To establish a more comprehensive indication of sandplain stability, a Victoria Plateau dune was dated using OSL. The Victoria Plateau dunes are partly degraded and stabilised by a heavy vegetation cover (where not disturbed by agricultural activity). These dunes are a striking element of the wider sandplain and provide clear evidence of a fundamentally different climate regime. In general, the dune fields are poorly organised; this is attributed partly to erosion, but it is also possible that the dune fields were never strongly organised. The dune was hand-augered to 8 m, revealing evidence for re-working of the upper 2 m between 18.4 ± 2.1 ka and 23.5 ± 3.5 ka, essentially coincident with the Last Glacial Maximum (K.-H. Wyrwoll et al., unpubl.). Prior to this, dune instability occurred at different times extending back to 221 ± 25 ka.

Additional OSL dating was undertaken on low dune forms on sandplains associated with the Gordon River, approximately 100 km north of Albany. Here, a range of channel deflation–source bordering aeolian deposits occurs on the eastern margins of the river (Fig. 2). These are a major contributor to the sandplain that lies further to the east, and which extends beyond the dune terrains. Large sand mounds have accumulated at the margins of the river channel, with 10s-of-metre-scale elevations evident along the river margins for some 30 km. These hummocky sand mounds grade into well-defined bed-forms, initially taking the form of poorly-defined parabolic forms, which further to the east grade into well-defined linear dune forms that constitute a large part of the sandplain. The OSL dates from these dunes range from 19.2 ± 0.9 ka to 25.8 ± 1.1 ka (K.-H. Wyrwoll *et al.*, unpubl.). A date of 55.6 ± 2 ka from the base of the dune provides the age of the substrate sandplain on which the later dune sediments were deposited. From the veracity of the dates obtained, it can be claimed with confidence that dune formation was coincident with the Last Glacial Maximum.

Figure 2. Gordon River sandplain with associated linear dune forms and dune section.

Both the Victoria Plateau and Gordon River dune dates attest to significant reworking and dune mobility, being especially emphasised by the Last Glacial Maximum dates. This was clearly a time of considerable hydrological stress for the inland regions, with the potential for significant reworking of the sandplains. Such reworking has clear pedological implications that need to be incorporated into any consideration of soil–vegetation associations.

DETAILS OF THE AGE STRUCTURE OF COASTAL SANDPLAINS

Overall considerations of the age of the coastal succession underpinning the vegetation can be placed in the context of global sea-level events over late Neogene to Quaternary time scales. The relevance of late Neogene events relates to the Yoganup Formation and the Bassendean Sands. The Yoganup Formation forms a relatively narrow rim at the foot of the Darling Escarpment, but further north and south it takes on more extensive dimensions in association with the Eneabba Scarp (north) and the Whicher Scarp (south). Unfortunately, the age of the Yoganup Formation is poorly constrained, but with the likelihood that it is older than Pliocene.

Two reliable chronological anchors are provided in the coastal sandplain succession: (i) the Holocene sequences, associated with the marine transgression at *ca.* 6–7 ka; and (ii) the Last Interglacial highstand (Marine Isotope Stage 5e, *ca.* 130 to 115 ka) that is pervasive throughout the coastal stratigraphy. There are no marine units between the Last Interglacial and the Holocene, but that is not to claim that some of the dune forms or substrates may not have been active during other times in the Late Pleistocene. The thermoluminescence dates of Price *et al.* (2001) suggest pervasive coastal dune activity in the wider Perth coastal region between 70–90 ka, but thermoluminescence dates are problematic and are difficult to evaluate without a clear understanding of their stratigraphic context.

While the age control on the Holocene transgression and the Last Interglacial succession is secure, the age control on the older Middle Pleistocene succession is problematic. In the Perth Basin these dunes form barrier complexes, often with a clear topographic succession. They have been separated convincingly on the basis of their heavy mineral assemblages (Bastian, 1996), yet various attempts at dating them have remained problematic (Hewgill *et al.*, 1983; Murray-Wallace & Kimber, 1989; Price *et al.*, 2001; Hearty & O'Leary, 2008). The succession requires a focused OSL dating campaign before any firm conclusion can be drawn as to the numerical age of these barrier complexes. On the basis of biostratigraphic evidence, a late Pliocene/early Pleistocene age has been proposed for the Bassendean Sands (Kendrick *et al.*, 1991). Less recognised, and less extensive, is the Yoganup Formation that lies at the foot of the Darling Escarpment, to which it is very difficult to ascribe an age.

The Perth Basin coastal succession provides a useful guide to the likely stratigraphic architecture of other regions of south-western Australia. However, with the sedimentology showing significant differences, especially when compared with the south coast, pedological differences must be expected. A firm conclusion that can be drawn is that extensive coastal sandplains became possible with the deposition of the Bassendean Sands during the Pliocene; an age estimate based on the claim that the formation is interpreted as the dune facies of the underlying marine Ascot Formation (see Kendrick *et al.*, 1991).

SOIL CHARACTERISTICS AND STATUS OF THE SANDPLAINS

The common attribute that characterises inland sandplain soils is their nutrient-poor status, with low concentrations of N, P, sulfur, base cations (calcium, magnesium, potassium) and micronutrients (copper, molybdenum, zinc) (*e.g.*, Bettenay, 1984; chapter 4). These infertile soils are associated with characteristic plant communities (Lambers *et al.*, 2010; chapter 4). Given the origins of many of the inland sandplains as a direct by-product of deep weathering events, the strength of this weathering inevitably leads to a low soil nutrient status. Profile development on sandplains is variable (Bettenay, 1984; McArthur, 1991) and can be subdued and restricted to the upper few decimetres of a profile, marked by the accumulation of organic matter and limited leaching. While an impression of general uniformity of sandplain soils attributes is apparent, there is significant variability in soil characteristics and their landscape position when considered in more detail.

Case study: landscape characteristics and soils of the Victoria Plateau sandplain

The work of Newsome (2000) demonstrated an essentially *in situ* origin for the Victoria sandplain, with sand provenance linked directly to the regional bedrock geology. In more detail, the Victoria sandplains relate to a range of soil-landscape systems, defined in terms of geomorphological and soil attributes. The soils of the Victoria Plateau were first mapped at small scale in the Atlas of Australia Soils (Northcote, 1960). Later, Beard (1976) based his vegetation map units in the region on soil catena relationships and McArthur (1991) provided detailed morphological descriptions and physical and chemical analyses of a number of reference soils at selected locations between Mingenew and Northampton. This work was followed by detailed mapping of the soil landscapes of the region (Rogers, 1996) which defined a number of systems and subsystems associated with the northern sandplain and the Victoria Plateau.

The systems identified all have occurrences of deep sand that are associated with kwongan vegetation. They overlie sedimentary lithologies associated with the Tumblagooda Sandstone at the northern extent of the sandplain, Permian sediments to the south and east, and Jurassic sediments in the west. The dominant systems typing the sandplain are Binnu North, Binnu East, Eradu and Casuarina.

(i) Victoria sandplain and associated land systems

a) Binnu (East and North) systems – east and northern Victoria Plateau: these systems have numerous dune ridges on alluvial valley slopes and are underlain by Permian sediments. The sands are up to 15 m deep and overlie laterite gravel that is indurated at depth. On the northern margin of the Victoria Plateau and the Kalbarri sandplain, the sands are associated with the Tumblagooda Sandstone.

b) Eradu system – southern Victoria Plateau: generally, level to gently undulating sandplain developed on the sedimentary rocks of the Yarragadee Formation (Jurassic). The surface sands (deep, yellow siliceous clayey sands and sandy earths, and minor occurrences of pale sands) are of variable thickness from 80 to > 800 cm, but typically 250 to 550 cm. The sands may be underlain by ferruginous gravels between 50 and > 150 cm deep which become cemented and dense with depth. The gravels overlie mottled clay that becomes bleached or pallid with depth.

c) Casuarina – southern Victoria Plateau: on the south-western margin of the Victoria Plateau the sandplain is dissected and consists predominantly of pale deep siliceous sand, which overlies a lateritic duricrust and gravels, which in turn are underlain by mottled and pallid clay.

(ii) Characteristics of the soils associated with the sandplain land systems

a) General soil characteristics:

Rogers (1996) identified six main soils series covering the inland sandplains: Eradu, Eurangoa, Indarra and Allanooka, Casuarina and Balline (*e.g.* Indarra Series; Fig. 5) These soils occur in areas described by Pate & Beard (1984) as being associated with a scrub-heath vegetation association. The Eradu, Eurangoa and Indarra series are yellow deep siliceous sands (Schoknecht & Pathan, 2013). Eradu is a loamy sand, transitioning to clayey sand that occurs extensively across the sandplain. The Eurangoa series, which occurs predominantly in the Binnu system in the north and east of the plateau, is lighter in texture than the Eradu series, with sands transitioning to clayey sand at depth. Colours of the Eurangoa series are usually paler than those of the Eradu series and are characterised by the presence of a conspicuously bleached A2 horizon, indicating that periodic waterlogging and lateral flow occurs in these soils (Northcote, 1983), despite the sandy textures. The Indarra series occurs on crests and rises of the Eradu system and the dunes of the Binnu systems. The Indarra is slightly lighter in texture (sand throughout) than either the Eradu or Eurangoa series.

The Allanooka, Balline and Casuarina soil series are all pale deep sands, most of which have a conspicuously bleached A2 horizon. The Allanooka series is characterised by a loam sand surface texture, pale yellow to yellow B2 horizon, and medium to coarse sands to 120 cm. They occur on long gentle slopes of undulating sandplain similar to the Eurangoa, but with less clay in the subsoil. The Balline series (loose grey coarse sand) occurs on the flat to gently undulating sandplain and as spillway sand below lateritic scarps, in the northwest of the plateau near the coast. This soil series is characterised by ~70% coarse sand (cf. 20% for Eurangoa and Casuarina series). The Casuarina series, which occurs in the southwest of the plateau is a loose sand over sandy gravel (30–70% laterite) at about 30 to 50 cm on to cemented laterite between 80 and ~120 cm that forms in association with the Eradu series.

b) Detailed textural characteristics of sandplain soils:

The textural characteristics of the surface grain-size distribution (< 2 mm) are generally similar for the Eradu, Eurangoa, Indarra and Casuarina soil series. The textural characteristics are of well-defined, well-sorted unimodal distributions (Fig. 3). Sorting is better defined for the Indarra series than for either the Eradu or Eurangoa series which is expected given the strong aeolian imprint on these sands. The Casuarina series contains more coarse-grained material than either of the yellow sands. It is similarly noteworthy that the Balline series is dominated by coarse sand, although formed on the sandplain.

While the surface textures of the respective soils series are very similar, considerable differences in textural characteristics occur with depth. These profile differences are well highlighted in the differences in clay content (Fig. 4). Both the Casuarina and Indarra series have less clay in the upper part of the profile compared with the Eradu and Eurangoa, but there is more clay at depth (about 120 cm), suggesting greater eluviation/illuviation in these profiles, and different soil water characteristics.

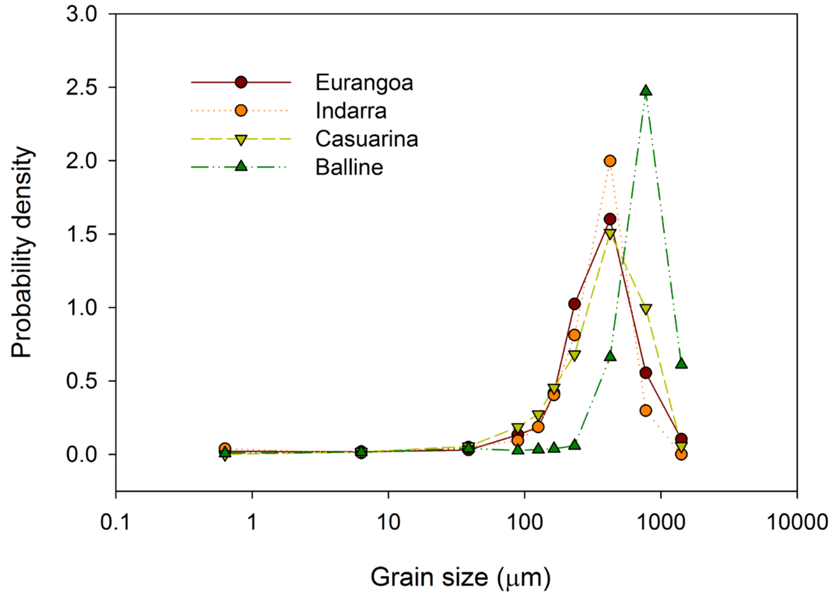

Figure 3. Comparison of grain size distribution (< 2000 μm) of the surface for some sandplain soils on the Victoria Plateau.

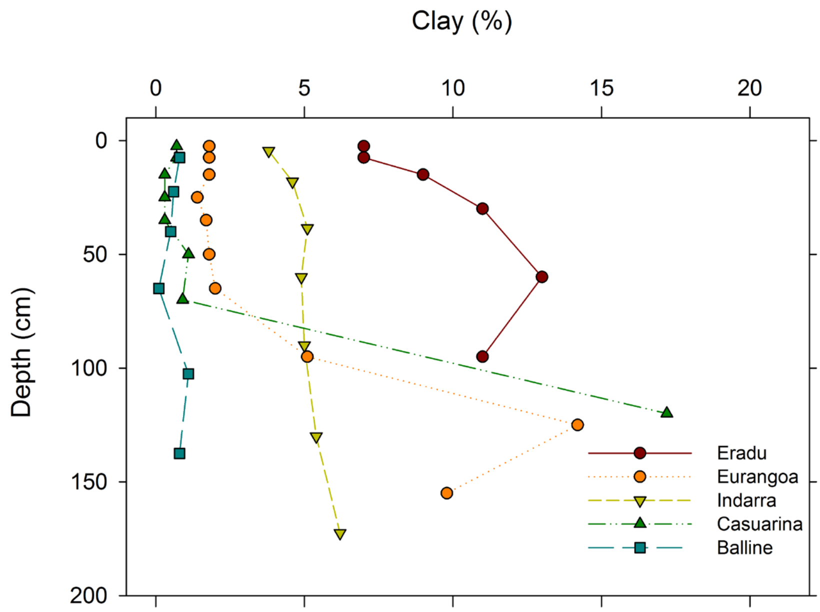

Figure 4. Comparison of clay content with depth for the deep sands on the Victoria plateau (data McArthur, 1991; Rogers, 1996).

c) Chemical attributes of Victoria Plateau sandplain soils:

Rogers (1996) and McArthur (1991) provided detailed chemical analyses of the Eradu, Indarra, Eurangoa, Casuarina and Balline series. Soil pH (H_2O) in the profiles can range from 5.3 to 6.8 or acidic to neutral (Hazelton & Murphy, 2007). The soils are low in salt (EC ranges of 1 to 5 mS $^{m-1}$) and aluminium (Al) (< 1 mg kg^{-1}), and in plant-available K (< 10 mg kg^{-1}), which may be limiting plant growth (chapter 4). Organic carbon concentrations are extremely low in the surface horizon (0.39 to 0.67%). High plant-available P concentrations (up to 25 mg kg^{-1}, extractable in bicarbonate) are associated with high organic carbon in the surface sands. One measure of the ability of a soil to retain P is the Phosphorus Retention Index or PRI (Allen & Jeffrey, 1990). This is a direct measure of P sorption and involves mixing a quantity of soil in solution with a single amount of P for a set period of time. Generally, minimum PRI values for the Victoria Plateau sandplain soils are in the range of 0.3 to 1.2 mL g^{-1} (very weakly absorbing or desorbing), except when clay content is high (and also likely associated with high sesquioxide levels) and PRI values can reach 33 mL g^{-1} (*i.e.* strongly absorbing). However, where the clay content is high in the profile (usually below 1 m) the plant-available P is < 2 mg kg^{-1}.

d) Comparison with the sands of the Yilgarn Craton sandplain:
While generally not as well expressed as inland sandplains developed on sedimentary rocks, the sandplains are extensive on the metamorphic and igneous terrains of the Yilgarn Craton (*e.g.*, Carroll, 1939). Details of the physical and chemical properties of the sandplain soils of the Yilgarn Craton are available in a number of publications (*e.g.*, Bettenay, 1984; McArthur, 1991; profiles KER 1 and 2 and MER 1; Grealish & Wagnon, 1995).

Clay content in the Yilgarn Craton sandplain soils can differ markedly compared with the Victoria Plateau sandplain soils. The latter tend to have a lower clay content, at least in the upper profile, with lighter textures (sands and loamy sands) compared with the Yilgarn soils, which commonly have loamy sands and sandy loams or heavier textures. However, chemical fertility is generally poor, regardless of source, on both the Victoria Plateau and Yilgarn Craton. The soil pH of the Victoria Plateau sands is usually less acidic than that of the Yilgarn soils. As the pH of the Yilgarn sandplain soils can be very low throughout the profile, Al toxicity may influence the vegetation. Victoria Plateau pH levels tend to be moderately acid to neutral throughout. Regardless of provenance, cation-exchange capacities are in the order of 1 or 2 $cmol_c$ kg^{-1}. Exchangeable cations (aluminium, calcium, magnesium, manganese, potassium and sodium) are very low to low, in the order of < 0.01 to 2.5 $cmol_c$ kg^{-1}. Organic carbon content (Walkley-Black) concentration in surface levels is typically < 0.5%, and likely to be 0.05% or less in the subsoil – these represent extremely low values by agricultural standards.

Sandplain soils are very low in the essential plant nutrients N, P and K (chapter 4). While total N varies seasonally, concentrations of about 0.3 mg N kg^{-1} soil have been measured in sandplain soils (McArthur, 1991; Lambers *et al.*, 2010). These levels are extremely low, resulting in C/N ratios of about 20. At this level, the organic matter breaks down less readily, and the small amount of N is far less accessible to plants. Levels of P and K are also low and, where the PRI is high (*i.e.* soils with high clay or sesquioxide concentration), plant-available P concentrations are still very low. Exchangeable K in the order 1 $cmol_c$ kg^{-1} or less have been measured in sandplain soils.

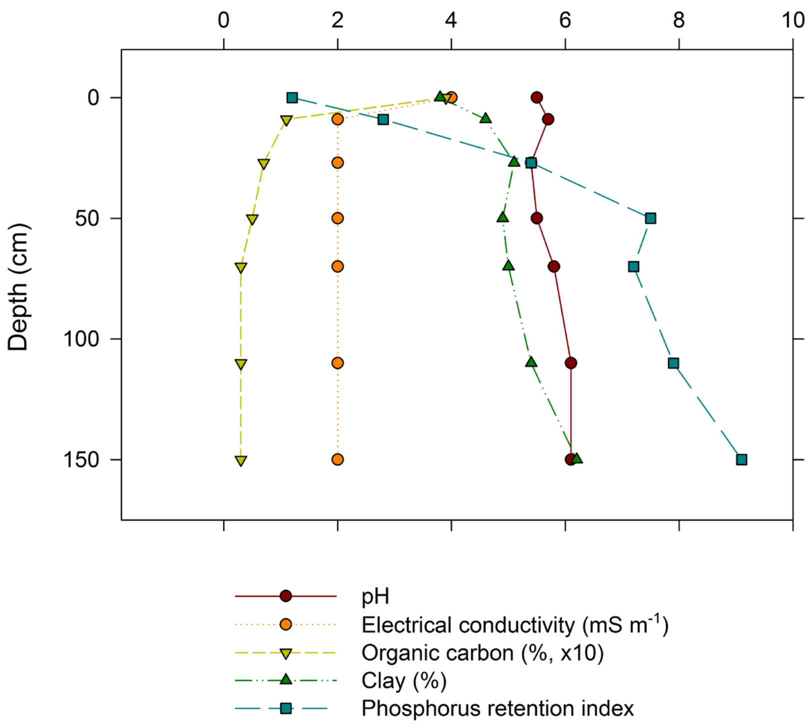

Figure 5. Changes in chemical properties with depth for Indarra soil series: pH H_2O – soil pH in water; EC – electrical conductivity; OC W/B – organic carbon (Walkey-Black method); PRI – Phosphorus-Retention Index (data Rogers, 1996).

SOIL EVOLUTION THROUGH TIME: A CHRONOSEQUENCE OF SOILS ON THE JURIEN BAY COASTAL SANDPLAINS

The sandplain succession of the Swan Coastal Plain extends back to at least the Pliocene and maybe even further (see above). The succession provides an important opportunity to study the development of soils and their associated plant communities over long time periods. Soils on the three dune systems form a chronosequence – a series of soils that differ in their formative properties only in the time since the onset of their development. This means that parent material, vegetation, climate and topography should all be constant (Jenny, 1941). Such long-term chronosequences are of scientific importance, because they provide a rare opportunity to understand ecosystem processes that cannot be addressed through conventional experimentation.

Detailed studies have been conducted in the vicinity of Jurien Bay, approximately 200 km north of Perth. The area has a Mediterranean climate, with cool moist winters and hot dry summers, and all three of the main dune systems (Quindalup, Spearwood, Bassendean) occur in this area, arranged as barrier complexes parallel to the coastline. This area is of particular interest given its status as a global biodiversity hotspot with particularly high levels of endemism (Hopper & Gioia, 2004). Examples of the three key stages in soil development are shown in Fig. 6. Similar soils are described from sites at Yalgorup National Park, near Waroona by McArthur (1991).

Soils on the youngest dune system (Quindalup dunes) consist of calcarenites (*i.e.* sand-sized carbonate grains), with carbonate contents of up to 80%. The carbonate weathers rapidly (in terms of soil development) and leaches from the profile, so that surface soil horizons on older Quindalup dunes (*i.e.*

after several thousand years of pedogenesis) can contain less than 20% carbonate. The Quindalup soils are light grey to greyish brown and exhibit a weakly developed surface horizon enriched with organic matter overlying several metres of unweathered calcareous sand (Fig. 6a). The young Quindalup soils are strongly alkaline (~ pH 9) and contain low concentrations of organic matter and N, but relatively high concentrations of P (> 300 mg P kg^{-1}).

Soils developed on the Spearwood succession are easily identified by their bright yellow subsurface horizons (Fig. 6b). The soils contain > 95% quartz sand, with the original carbonate matrix having been removed completely by leaching during at least 100,000 years of pedogenesis. The yellow colour is derived from iron-oxide coatings on quartz grains. The young Spearwood soils are underlain by Tamala Limestone, often capped by a calcrete, at shallow depth (often < 1 m below the surface), whereas older Spearwood soils are many metres deep. In addition to the absence of carbonates, soils on the Spearwood dunes are characterised by a neutral to slightly alkaline pH and much lower concentrations of total P (typically 10–30 mg P kg^{-1}) and base cations compared to the younger Quindalup soils. Older Spearwood soils (*i.e.* formed prior to the Last Interglacial Maximum) exhibit an incipient bleached light grey horizon near the surface, from which iron oxides have been leached – an indication that these older soils are transitioning towards the final stage of soil development along the chronosequence.

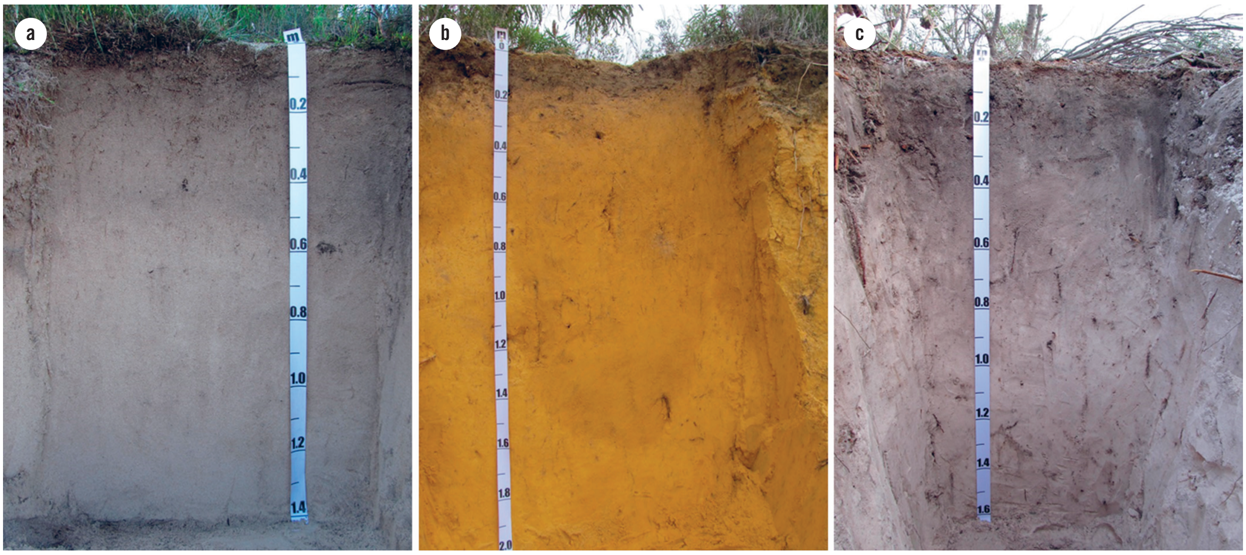

Figure 6. Soil profiles representing three of the major stages of soil development along the Jurien Bay chronosequence, Western Australia. (a) Young soil (~1000 years old) developed in carbonate-rich Quindalup dunes (~70% $CaCO_3$), showing weak horizon development. The pH is alkaline and there is little organic matter, but relatively high concentrations of total phosphorus (P) (300–400 mg P kg^{-1} in surface horizons). The soil contains around 98% sand-sized material. The soil is a Rudosol in the Australian classification system and an Entisol (Typic Xeropsamments) in the US Soil Taxonomy system. (b) A soil profile (~100,000 years old) developed on Spearwood sand. The profile has been leached of carbonates, leaving residual quartz sand stained yellow by iron oxides. The profile shows a shallow, organically-enriched A horizon at the surface, and a deep yellow B horizon containing > 96% quartz sand extending > 4 m to limestone. The pH is neutral, and total P concentrations are very low (~10 mg P kg^{-1} in surface horizon). Despite its age, the soil is an Entisol (Xeric Quartzipsamments) in the US Soil Taxonomy system. It qualifies as a Podosol in the Australian soil classification system. (c) A very old soil (~1,000,000 year old) developed on Bassendean sand. The iron oxides present in the Spearwood sand profile have been leached from the profile as humus-iron complexes, leaving a deep bleached elluvial (E) horizon that contains virtually no nutrients. The leached iron-containing compounds form a deep 'coffee rock' horizon in some locations. The pH is slightly acidic, and total P concentrations are extremely low (< 10 mg P kg^{-1}). The E horizon contains 99% quartz sand. Despite its age, the soil is an Entisol (Xeric Quartzipsamments) in the US Soil Taxonomy system. It qualifies as Podosol in the Australian soil classification system.

The final stage of the chronosequence is represented by soils developed on the Bassendean Sands (Fig. 6c). This formation was deposited during the early Pleistocene–Late Pliocene, with the sediments having an initial lower carbonate status the Spearwood and Quindalup associations (discussed above). Despite this, the soils of the Bassendean Sands are taken to represent the extremely infertile end-point of an exceptionally strong nutrient gradient (Laliberté *et al.*, 2013a). The iron oxides that provide the yellow colours of the Spearwood sands have been completely leached from the profile, through the formation of complexes between iron and organic matter. This leaves clean quartz grains forming bleached white eluvial (E) horizons that can be more than 3 m thick. In some locations, the iron–organic matter complexes are deposited in spodic horizons in the deep subsoil, forming a brown (iron–humus) material known locally as coffee rock. Soils on the Bassendean Sands are slightly acidic and contain extremely low concentrations of total P (< 10 mg P kg^{-1}). They therefore constitute some of the lowest-P soils known globally.

Only a few locations in the world contain long-term soil chronosequences. Well-known examples include the Hawaiian Islands chronosequence (Crews *et al.*, 1995), the Franz Josef chronosequence in New Zealand (Walker & Syers, 1976), and the Cooloola chronosequence in eastern Australia (Thompson, 1981). In all these sequences, soils show consistent patterns of nutrient dynamics over long time scales, as identified by Walker and Syers based on studies of a series of chronosequences in New Zealand (Walker & Syers, 1976). In the Walker and Syers model, young soils contain little N, because this element is absent in most parent materials, including the coastal sands in south-western Australia. However, N accumulates rapidly in the early stages of pedogenesis through biological nitrogen fixation (*e.g.*, Menge & Hedin, 2009). In contrast, P concentrations are relatively high in young soils, but decline over time as P is lost by leaching and erosion at a greater rate than it is replenished by bedrock weathering and atmospheric deposition.

These changes have profound ecological consequences, because the productivity of plant communities tends to be limited by the availability of N on young soils, but becomes increasingly limited by P availability as ecosystems age. This eventually leads to a decline in plant productivity and biomass on old, strongly-weathered soils (Wardle *et al.*, 2004). There are also marked changes in the composition of plant communities as ecosystems age, because increasing infertility favours slow-growing species with specialised mechanisms to acquire and retain P. Interestingly, vascular plant diversity also increases as soils age, exemplified by changes along the Jurien Bay chronosequence (Hayes *et al.*, 2014; Laliberté *et al.*, 2013a).

The Jurien Bay chronosequence shows patterns of soil nutrients that correspond to the Walker & Syers (1976) model of nutrient transformations during pedogenesis (Laliberté *et al.*, 2013a; 2013b). In particular, soil phosphorus declines from the youngest to the oldest soils, leading to strong P limitation of plant growth on the oldest soils (Laliberté *et al.*, 2012). In contrast, plant growth is likely limited by N, K or micronutrients on the youngest calcareous soils. Phosphate that is considered most readily available for biological uptake (‘plant-available phosphate’) ranges from 2–3 mg P kg^{-1} on the young calcareous dunes to < 0.5 mg P kg^{-1} on the older dunes (determined by extraction with anion-exchange membranes). Effective cation exchange capacity is also very low on older dunes – it is dominated by calcium in all dunes, with extremely low concentrations of K throughout (< 0.1 cmol$_c$ kg^{-1}). Cation-exchange sites are predominantly on organic matter on the oldest soils.

At present, the numerical time estimates necessary to provide a secure time base for these pedological changes along the coastal sandplain chronosequences are not available. We know from the general stratigraphic expression of the Tamala Limestone succession that this involves the Last Interglacial and earlier Middle Pleistocene interglacial stages. The Bassendean Sands provide an early Pleistocene–late Pliocene estimate for the earliest events. The degree to which the age of soil profiles corresponds to these major stratigraphic divisions remains to be established. An OSL dating program, presently underway, will provide at least part of the answer to this question.

CONCLUDING DISCUSSION

The association of inland sandplains with deep weathering, and the detailed geomorphological characteristics and palaeoclimate requirements, point to the possibility that the south-western Australian inland sandplains have a history that may only extend into the late Miocene–early Pliocene. It is clear that the inland sandplains have been extensively reworked, albeit at a local scale, during Pleistocene climate events. So, at least from a Quaternary perspective, the stratigraphy and age structure of inland sandplains indicates that they cannot be considered to have been climatically 'buffered', but have in fact been imprinted repeatedly by past climate events. Taken together, these inferences challenge elements of both the OCBIL theory and the more recent claims that species richness of the Southwest Australian Floristic Region is primarily the product of a relatively stable Pleistocene climate (Sniderman *et al.*, 2013). However, while Pleistocene climates of the inland regions were far from stable, the same cannot be claimed for some of the coastal rim regions of the southwest.

The existence of the karri–tingle forest of the southwest is linked strongly to the high annual precipitation and edaphic controls specific to the region (*e.g.*, Christensen, 1992). When these distinctive forest associations are considered in the context of their specific requirements and the more general geographical setting of the wider region, it is apparent that there were no opportunities for the vegetation association to migrate when hydrological requirements were not satisfied and the extreme southwest was essentially encased by more arid climates. Even the land extensions resulting from the lower sea level of glacial stages would have been insufficient to provide necessary refugia or migration opportunities, although this does not preclude restricted contractions and expansions of local ranges if moisture conditions changed.

Pickett (1997) showed from lake sediments and their associated pollen sequences from the Perth region that a reduction in precipitation took place at the Last Glacial Maximum, but found no indication of a significant collapse of the winter (westerly controlled) precipitation regime. This, despite the fact that some 130 km further east the Avon River presented a broad sand-based channel, without a significant riparian rim which was extensively deflated both during the summer and winter months (Wyrwoll, unpubl.). This is consistent with dry conditions prevailing for much of the year. Furthermore, in an early global circulation model experiment, storm tracks embedded in the westerly (winter) rainfall regime of south-western Australia essentially followed their present paths (Wyrwoll *et al.*, 2000). This has been generally supported by more recent modelling results (Rojas *et al.*, 2009; Sniderman *et al.*, 2013), despite the complexity of responses across a range of global circulation models.

Overall, a model emerges whereby at glacial maxima, inland sandplain environments were strongly stressed with limited vegetation cover and general erosional instability, while coastal rim sandplains experienced a hydrological environment not that much different from that of today, with general geomorphological stability. The model points to a strong regional hydrological gradient at the Last Glacial Maximum, a claim supported clearly by the Gordon River succession (see earlier). For such extensive dune migration to occur, the vegetation cover at the Last Glacial Maximum must have been impoverished. It is remarkable that karri forest may have prevailed at the same time some 150 to 200 km to the southwest (Wyrwoll *et al.*, 2000).

A model in which inland sandplain regions experienced a significant change of climate to much drier conditions, with coastal margins retaining much of the present rainfall amounts and water-balance characteristics, may require modifications of the claims of Sniderman *et al.* (2013) and Hopper (2009). Instead of either (i) a general climatically-buffered sandplain setting, or (ii) Pleistocene climate stability, the revised model indicates that major climate shifts impacted the inland sandplains, while elements of the coastal rim remained buffered. This model of sandplain origin and stability is highly speculative, but the development of cosmogenic isotope and OSL dating techniques, linked to detailed sediment studies, would remove much of the uncertainty. Despite these uncertainties, it is clear that the inland sandplains are not simply stable, long-term remnants in the landscape – there has clearly been reworking and the refreshing of pedological imprints.

The inland sandplains, in their present expression, may be no older than late Miocene–early Pliocene, with the present-type inland sandplain environments not coming fully into existence until possibly the latest Pliocene–early Pleistocene. However, this does not preclude the existence of prior sandplain terrains. Given the stability and low relief characteristics of both cratons and sedimentary basins, weathering 'residuals' analogous to present-day sandplains, are likely to have been features of these regions throughout much of their geological history. Present-day sandplains are simply the expression of the most recent (Paleogene/Neogene) erosional cycle. In this way, the apparent contradiction between the plant–fossil phylogenetic and geomorphological evidence can be reconciled.

The presently available stratigraphic evidence points to extensive reworking of inland sandplains during periods of extreme hydrological stress that occurred at various times in the Pleistocene, with glacial maxima being especially important. Using the Last Glacial Maximum as a guide, we propose a hydrological model in which the southwest climate was characterised by strong coastal to inland precipitation gradients, with extreme hydrological stress prevailing in inland areas. At the same time, much of the coastal rim of south-western Australia retained relatively benign hydrological states, hence experiencing a more stable long-term hydroclimate environment. This model recognises the potential long-term stability of coastal sandplains, compared with the extensive restructuring of the inland sandplains by erosional and transport processes.

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