Hidden in the sands of time:
geoarchaeology of sandstone landscapes
in the Keep River region, Northern Territory, Australia

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CHAPTER FOUR

*In Situ* Cosmogenic Isotope Dating of Sandstone Bedrock and Sand Sheets

*As the physical processes shaping the land can be seen to unfold over hundreds and thousands of years, so can the .. processes which are evident in the archaeological data.*  
– Gosden and Head (1994: 113)

4.1 Introduction

This chapter aims to outline and model the broader geomorphic processes of landscape evolution in the Keep River region, including quantifying rates of plateau denudation using *in situ*-produced cosmogenic radionuclides, $^{10}$Be and $^{26}$Al. The denudation history of the plateaux is complemented with the depositional chronology of the genetically linked sand sheets determined from profiles of $^{10}$Be and $^{26}$Al concentrations. The innovative use and geoarchaeological application of cosmogenic radionuclides to measure sediment accumulation involves some experimentation; hence the theoretical background and accumulation models are outlined below. The long-term sand sheet chronology is subsequently compared to the chronology determined by luminescence dating techniques in Chapter Six.

4.2 *In Situ* Cosmogenic Isotope Dating

As indicated in Chapter Two (refer 2.4.1), the accumulation of cosmic radionuclides can be utilised as radiometric clocks to elucidate the erosion rate ($\varepsilon$) of geomorphic formations or exposure age of a previously unexposed surface ($t_{\text{exp}}$). A decade ago, Lal (1991) presented the first definitive publication for modelling exposure dating and defining the cosmic radionuclide (CRN) method. The general theory and model equations are outlined here in increasing order of complexity; first for exposure age dating, second for burial dating, and finally, sediment accumulation. The latter is the major focus of this chapter.
4.2.1 Exposure Dating - Theory and Equations

The cosmogenic radionuclides $^{26}\text{Al}$ ($T_{1/2} = 0.71$ Ma) and $^{10}\text{Be}$ ($T_{1/2} = 1.5$ Ma) are produced by reactions of two main types of cosmic-ray secondary particles, nucleons and muons, with silicon and oxygen target atoms in quartz (Tuniz et al., 1998). The accumulation of cosmogenic radionuclides in the Earth’s lithosphere occurs as a function of time and depth below surface, and as such, is sensitive to the denudation rate of that surface. Only about a million atoms of cosmogenic radioisotopes are produced \textit{in situ} within a ten thousand year exposure period per gram of surface rock, hence the technique of Accelerator Mass Spectrometry (AMS) is needed to measure this signal (Tuniz et al., 1998). An exposure history refers to an estimate of the length of time ($t$ in years) the rock was exposed to cosmic rays and the average erosion rate ($\varepsilon$ in centimetres per year) experienced during exposure. On the assumption of an ideal surface undergoing zero erosion, the build-up in concentration of \textit{in-situ} produced radionuclides can be expressed by the following equation:

$$C_0 = P \left(1-e^{-\lambda t}\right) / \lambda$$

Where $C_0 =$ concentration (atoms per gram of SiO$_2$) at the present rock surface, $\lambda =$ radioisotope decay constant ($= \ln 2 / t_{1/2}$ years$^{-1}$), and $P =$ production rate at the sample site (atoms per gram of SiO$_2$ per year) as a function of latitude, $\theta$, and altitude, $A$.

For a constant production rate, the concentration ($C$) of radionuclides increases with time during cosmic-ray exposure, at first linearly and then more slowly until, after a few half-lives, the system reaches a secular equilibrium in which production equals decay. By knowing the rate of production ($P$) and measuring the surface concentration it is possible to estimate a surface exposure age from:

$$T_{\text{exp}} = - \ln \left(1- \frac{\lambda C_0}{P}\right) / \lambda$$
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Assuming zero erosion, model exposure ages are necessarily minimum ages and for younger surfaces (< 20 ka), denudation rates effectively have little influence. For surfaces older than many tens of thousands of years, denudation becomes increasingly important. In the presence of a constant erosion rate, given by \( \epsilon \), equation (1) becomes (Lal, 1991):

\[
C = P \left(1 - e^{-\left(\lambda + \epsilon \rho / \Lambda \right)t}\right) / \left(\lambda + \epsilon \rho / \Lambda \right)
\]  (3)

Where \( C \) is the measured surface radioisotope value at time \( t \),
\( \rho = \) density of the material (grams per cubic centimetre), and
\( \Lambda = \) attenuation length of cosmic-rays.

The attenuation length is defined as the thickness of material (i.e. rock) which is required to reduce the intensity of cosmic ray flux to a value which is \( 1/e \) (~ 0.4) of the surface (or original) intensity. The depth profile of the flux is given by \( P(x) = P(x=0) e^{-x/\Lambda} \) and hence \( \Lambda \) determines the effective penetration depth of active cosmic rays, which for the average rock density of 3 g.cm\(^{-3}\) represents a distance of ~ 50 cm depth (or equivalently ~ 150 g.cm\(^{-2}\)). In effect the denominator term \( (\lambda + \epsilon \rho / \Lambda) \) can be interpreted as the residency time radionuclides within the first 50 cm of the rock surface. The time required to remove a thickness of rock equivalent to the mean attenuation length \( \Lambda \) of cosmic rays is equivalent to the mean effective or apparent exposure age (\( t_{\text{app}} \)). By comparing equations (1) and (3), it is possible to define an apparent age, \( t_{\text{app}} \), which essentially incorporates the surface denudation rate and radionuclide decay to give an “effective half-life”:

\[
t_{\text{app}} = 1/\left(\lambda + \epsilon \rho / \Lambda\right) = 1/\lambda_{\text{app}}
\]  (4)

For sufficiently long exposures, cosmogenic concentrations saturate; at which point production is balanced by the sum of decay and removal by erosion. Thus for very small \( \epsilon \), where \( t >> t_{1/2} \), an expression for the denudation rate can be obtained, as follows:
\[ \epsilon = \Lambda / \rho \left( P_0 / C_s - \lambda \right) \]  

Under these conditions, the maximum saturation concentration at the rock surface is dictated by the magnitude of the denudation rate. The larger the value of \( \epsilon \), the smaller the allowed saturation value. Increasing denudation rates imply that a surface has been uncovered rapidly, and correspondingly surface saturation concentrations are decreasing. For the surface to achieve erosional equilibrium, the model erosion rate is necessarily a maximum one (\( \epsilon_{\text{max}} \)) (Nishiizumi et al., 1991) which alternatively corresponds with a model interpretation of minimum exposure age.

Using two radionuclides with different half-lives in principle allows a better simultaneous estimate of the denudation rate and exposure age variables. A plot of the concentration ratio of \(^{26}\text{Al}/^{10}\text{Be}\) against \(^{10}\text{Be}\), shows the limiting curves for denudation and exposure age, which define the steady-state erosion island (Fig. 4.1) (Lal, 1991). Theoretically, samples that plot on the upper curve (or the minimum exposure line) have undergone no burial or erosion (\( \epsilon = 0 \)) during the current or finite exposure, the duration of which is indicated by the distance along the curve (Fig. 4.1). Samples with ratios plotting on the lower curve (or maximum erosion line) represent erosional equilibrium conditions of exposure (\( t = \infty \)) (Nishiizumi et al., 1991). Samples plotting within the erosion island theoretically represent a unique and quantifiable combination of denudation and exposure but in practice, resolution (defined in terms of cumulative errors, technique sensitivity, precision, etc.) limits quantification to minimum exposure (\( t_{\text{min}} \)) and maximum erosion (\( E_{\text{max}} \)). Samples plotting below the steady-state erosion island would indicate a more complicated exposure history of burial (see 4.2.2) (Fig. 4.1), and in the absence of independent information, a minimum exposure is assumed (Gosse and Phillips, 2001). Samples plotting above the erosion island indicate measurement error (or sample contamination), except in the case where a sample received prior exposure at a site where the nuclide production rate was higher, such as through uplift (Lal, 1991; Brook, 1995).
Theoretically calculated variation in the ratios of $^{10}$Be and $^{26}$Al concentration as a function of $^{10}$Be concentration for unit nuclide production rates (atom/g/yr) (modified from Lal, 1991). Surfaces with different erosion rates plot in the shaded area, termed the steady-state erosion island. Lines running diagonally represent lines of burial such that with sufficient burial there is zero production ($P = 0$). Samples which have undergone cyclic periods of burial and exposure fall below the curve in the complex exposure zone.

Although the removal and radioactive decay of $in-situ$ cosmogenic radionuclides is well understood, the precision and accuracy of exposure histories depends primarily on the accuracy to which the rate of production of radionuclides can be determined. Production rates of $in-situ$ cosmogenic radionuclides are measured from sites whose exposure age is independently known (Nishiizumi et al., 1989; Clark et al., 1995). Radiocarbon dating is typically used for age control and thus production rate calibration is usually not available past ~20 ka, or for simplicity is otherwise assumed to have remained constant over the whole period of exposure (Nishiizumi et al., 1989; Bierman et al., 1996; Tuniz et al., 1998). Another method is to determine production rates experimentally, by exposing target materials to cosmic radiation at high latitudes for a number of years (e.g. Nishiizumi et al., 1996; Brown et al., 2000).
The cosmic ray flux changes as a function of latitude and altitude (Lal, 1991; Nishiizumi et al., 1991). Hence scaling factors are needed to obtain a relative cosmic ray flux for unknown sample locations. Using the known production rate at calibration sites it is possible to scale production to the location of an unknown latitude and altitude (Lal, 1991; Dunai, 2000; Stone, 2000). Production rates are not time independent and thus scaling factors have also been developed to account for geomagnetic field intensity variations and magnetic pole wanderings (Clarke et al. 1995; Dunai, 2000; Masarik et al., 2001). In addition, it is also necessary to correct for site specific effects such as geometric shielding of cosmic rays by horizon obstruction and attenuation on sloped surfaces (Dunne et al., 1999). The effects of topography and shielding by local vegetation are site specific, and for the dry savannah of the Keep River region these effects are negligible. In this study, the production rates from Stone (2001) are used. Specifically these are 5.1 at/g/yr for $^{10}$Be and 30.1 at/g/yr for $^{26}$Al at sea-level (SL) and high latitude (HL), with quoted errors of 6 – 8 %. No additional uncertainty is included, and no correction is made for geometric field variation. A full review of production rates for $^{10}$Be and $^{26}$Al is given by Kubik et al. (1998) and Gosse and Phillips (2001).

Whilst in situ cosmogenic isotope dating has the potential for solving many geomorphological and archaeological problems, it is important to acknowledge that the measured radioisotope concentration does not necessarily define a geological age. For recent or stochastic events, where correction factors are minimal, cosmogenic dating can provide an age effectively equivalent to a true age. For longer term processes of evolving landscape surfaces, cosmogenic dating provides an erosion rate which can be used to estimate the mean apparent age. Depth profiles of the concentrations of radionuclides, ratios of different radionuclides in the rock, and concentrations of radionuclides in the sediments produced by erosion can all provide measures of the rate of erosion (Harbor, 1999). The challenge is to correctly interpret the model results according to the geomorphic and geoarchaeological setting (e.g. Fig. 4.2).
Possible exposure histories for rocks sampled for in situ cosmogenic dating studies (modified from Dorn and Phillips, 1991), including (a) multiple cycles of exposure through erosion and burial, (b) constant centimetre-scale erosion of exposed landforms through physical weathering processes, (c) episodic large-scale mass wasting of exposed landforms, and (d) human abrasion of exposed rocks, such as engraving or tool making.

The measurement of more than one radioisotope offers the potential to differentiate phases of combined geomorphological processes, because each radioisotope has a half-life that dictates the time scale of its sensitivity as a chronometer (Lal, 1991). From in situ cosmogenic nuclide concentrations, exposure ages may be determined in the order of 5 ka – 5 Ma, and erosion rates in the order of ~0.1 mm to 100 mm.ka\(^{-1}\). Whilst the upper age limit is set by radionuclide saturation, the lower age limit is set more by the practical limitations of analysing kilograms of material rather than the present sensitivity of AMS.

### 4.2.2 Burial Dating – Theory and Equations

Burial dating was first proposed by Lal and Arnold (1985). It is similar to surface exposure dating in that both rely on the in situ production of cosmogenic radionuclides that necessarily decrease with depth below surface. However, the methods differ both in how they measure time intervals and in the problems they are useful for (Granger and Muzikar, 2001). In contrast to surface exposure dating, which measures accumulation from time zero (or event time), burial dating measures the relative decay of a reserve of radioactive radionuclides in buried sediments. The most straightforward case of burial dating assumes that material, whose initial radionuclide concentration is known, has been rapidly buried (e.g. washed into a deep cave, or covered by several metres of overburden), so that cosmogenic production effectively ceases (Granger and Muzikar, 2001). The
situations in which production does not completely cease upon burial, and which is the main subject of this chapter, is considered below (refer 4.2.3).

$^{26}\text{Al}$ and $^{10}\text{Be}$ are particularly useful for burial dating because the ratio of their production rates is well known, and roughly independent of factors such as latitude, altitude, and depth (Granger and Muzikar, 2001). As $^{26}\text{Al}$ decays more rapidly than $^{10}\text{Be}$, the $^{26}\text{Al}/^{10}\text{Be}$ ratio decreases exponentially with burial time. Pre-burial concentrations and hence erosion rates can be inferred from simple models of cosmogenic nuclide accumulation at the surface. This study also makes use of plateau denudation rates from \textit{in situ} cosmogenic dating. The age range of burial dating for $^{10}\text{Be}$ and $^{26}\text{Al}$ in quartz is in the order of 200 ka to 5 Ma (Granger and Muzikar, 2001), which accords within the depositional age provided by luminescence dating of the sand sheet sediments surrounding the Jinmium site (Fullagar et al., 1996).

\section*{4.2.3 Modelling Accumulation (Partially Shielded Burial)}

The model of \textit{in situ} cosmogenic isotope depth profiles follows the methods of Lal et al. (1987) for determining accumulation rates, $S$ (mm.ka$^{-1}$) in ice cores from Antarctica. For modelling purposes it is assumed that sediment accumulation has been constant for long periods, and has not been instantaneously or repeatedly buried (e.g. migrating sand dunes). For accumulating sediment sequences, production does not cease upon burial, and is referred to here as partially ‘shielded burial’. For an accumulating profile, or for partial burial of surface, the cosmogenic nuclide concentrations may increase as long as the buried surface remains shallow enough (i.e. low accumulation rates) that production outweighs decay (Braucher et al., 1998; 2000), after which the shape of the profile is dependent on the accumulation rate. Hence it is possible to model theoretical patterns of nuclide depth profiles assuming different initial concentrations and transport of source material and re-deposition or accumulation. This study considers an accumulating surface where deposition outweighs removal.
Under constant accumulation, depth \((X, \text{cm})\) can be used as a proxy for time (years), and the concentration \((C_{\text{TOT}}, \text{atoms per gram})\) as a function of depth is given by the following equation:

\[
C_{\text{TOT}}(x) = C_0(E_0)e^{\lambda x/S} + P_0(e^{\lambda x/S} - e^{-\rho x/\Lambda})/ (\rho S/ \Lambda - \lambda) \quad (\text{Equation 6})
\]

Where
- \(C_0\) is the nuclide concentration (atoms per gram) of the transported sediment (i.e. \(x = 0\)),
- \(E_0\) is the erosion rate of the contributing bedrock source (mm per year),
- \(P_0\) is the nuclide production rate (atoms per gram per year) at \(x = 0\),
- \(\rho\) is the density (grams per cubic centimetre),
- \(\Lambda\) is the cosmic ray absorption mean free path (grams per square centimetre),
- \(\lambda\) is the decay constant of the nuclide (units of inverse years), and
- \(S\) is the sediment accumulation rate (centimetres per year).

The initial surface concentrations of \(^{10}\text{Be}\) and \(^{26}\text{Al}\) in deposited sand are assumed to be related to the average erosional equilibrium concentration, \(C_0\), of the bedrock source (Fink et al., 2000). In this study, the value of \(C_0\) is given by the average CRN concentrations, \(C_i\), of all the bedrock plateaux samples. For any given value of \(C_i\) it is possible to estimate a corresponding maximum model erosion rate, \(E_i\), according to equation (7):

\[
C_i = P_i(0)/ (\lambda + \rho E_i / \Lambda) \quad (\text{Equation 7})
\]

This equivalence assumes that all the eroding bedrock surfaces contribute equally to the accumulation of sediment in a particular region, and here specifically the Keep River region. The mean erosion rate can then be determined from the average of all the \(E_i\).

Modelling of \textit{in situ} cosmogenic depth profiles can be determined using equation (6), outlined above, for a range of erosional equilibrium concentrations \((C_0)\) of source material and estimated
accumulation rates (S) for both \(^{10}\)Be and \(^{26}\)Al. For zero accumulation (including autochthonous weathering), samples from any given depth will have maintained their relative shielding per unit time (i.e. the surface level remains constant) and the cosmogenic profile should appear similar to a bedrock core with close to exponential decay. On a log scale, nuclide depth profiles show steadily decreasing nuclide concentrations (Fig. 4.3a).

For a depositional profile where the sediment accumulation rate (S) is constant, the picture is more complex. It is assumed that the bedrock erosion rate, characterised by \(C_0\) (or \(E_0\)) is the only source of transported sediment. With decreasing bedrock erosion rate (increasing initial radionuclide concentrations), the shape of the nuclide depth profiles become more parallel (Fig. 4.3b). For the case where the initial radionuclide concentration of transported sediment is constant, the radionuclide depth profiles typically decrease with depth and become more flat with increasing sedimentation rate (Fig. 4.3c). The aim is to fit an envelope of modelled profiles to the measured nuclide concentrations from a sand sheet profile, whereby the point where \(x = 0\) is fixed by the initial erosional equilibrium concentration (\(C_0\)) and different values of sedimentation rate (S) are used to match the shape. The complexities of radionuclide inheritance, accumulation, erosion and decay mean that it is better to focus on the entire profile rather than individual points (Brown and Bourles, 2002).

In this way the evolution of \(^{10}\)Be or \(^{26}\)Al concentrations as a function of depth for a profile can allow quantification of sedimentation (S) and erosion (E). Moreover it can differentiate various patterns of erosion and burial (Phillips, 2000), including that resulting from in-situ (autochthonous) conversion of bedrock to soil or regolith and constant accumulation of transported (allochthonous) sediments (Brown et al., 1998; Braucher et al., 1998; 2000; Heimsath et al., 1999). Examples of burial dating include the origin of stone-lines in lateritic environments (Braucher et al., 1998, 2000; Brown et al., 1994), and the evolution of desert sand-dunes and sand plains in arid Australia (Fink et al., 2000). Further details and synopsis of burial dating can be found in Lal (1991), Phillips, (2000) and Granger and Muzikar (2001).
Figure 4.3 Theoretical patterns of nuclide depth profiles assuming (a) zero sedimentation ($C_{\infty=0}$), (b) varying erosion rate, and (c) varying sedimentation rate (modified from Fink et al., 2000; Phillips, 2000). The model curves assume values for production of $^{10}$Be and $^{26}$Al scaled to the latitude (15.43 °S) and altitude (10 m ASL) of the Keep River region. Refer text for explanation.
Bioturbation buffers the average cosmogenic nuclide concentrations in surface soils against short-term changes in erosion (Brown et al., 1995), hence the effects of bioturbation may need to be considered in the determination of burial rates of sand sheet sediments. In theory, bioturbation will homogenise the nuclide concentration throughout its region of activity to a value similar to that of a surface unaffected by bioturbation, and hence can be used to differentiate the bioturbated zone from lower layers unaffected by bioturbation (Brown et al., 1995; Bracher et al., 2000). In reality all surfaces are affected by bioturbation, but the ideal profile of the model remains valid. In the Keep River region, the greatest bioturbating influence is likely to come from prevalent termite activity, with rates of topsoil reworking estimated from 25 mm.ka\(^{-1}\) (Holt et al., 1980) to 200 mm.ka\(^{-1}\) (Hart, 1995). From studies of stoneline profiles in southwest Australia, Colin et al. (2001) derive similar rates of termite excavation in the order of 20 mm.ka\(^{-1}\) using \textit{in situ} produced \(^{10}\)Be. Termite activity is generally restricted to above the water table, which in a monsoonal environment may be extremely variable. Further consideration of the effects of bioturbation on stratigraphic integrity is given in Chapter Six in relation to luminescence dating, and any microstratigraphic evidence of bioturbation is investigated Chapter Five.

The two main types of cosmic-ray particles, nucleons and muons, have different attenuation lengths. The effective attenuation length of neutrons is short (\(\sim 150 \text{ g cm}^{-2}\)) relative to that of muons (\(\sim 750 \text{ g cm}^{-2}\)) (Brown et al., 1995; Granger and Smith, 2000). This effectively means that neutron-induced production dominates near the surface whilst muon-induced production only becomes significant for solid bedrock profiles > 10 m (Fig. 4.4). Samples taken in the Keep River region are neither deep enough nor old enough (refer 4.3.1) for muon contribution to be significant.
4.2.4 *In situ* Cosmogenic Dating and Sand Sheet Formation

Sand terrains comprising ergs, dunes and sand sheets occur in the arid and semi-arid and monsoonal environments of the Sahara, Arabia, central Asia, southern Africa and Australia, yet the origin of many of these sand terrains remains uncertain (Newsome, 2000). In western Arnhem Land, it has been argued that sand deriving from the nearby sandstone plateau began accumulating at the same time as the initiation of human occupation around 50 - 60 ky BP (Roberts et al., 1994b), whether as a result of human firing practices (e.g. Hope et al., 1985; Jones, 1985) or as a response to climate (e.g. Ross et al., 1992). There is some question over the justification of the coincidence of commencement of occupation and sediment accumulation (refer 2.5.2). Nevertheless, these observations imply that the sand sheet sediments have accumulated over the top of an exposed pediment. Further evidence is required to differentiate the evolution history of the sand sheets from a number of possible options, including: (i) the *in situ* (autochthonous) conversion of the underlying
saprolite (Mory and Beere, 1988), (ii) long-distance transport of aeolian sediments (Pye and Gardner, 1981; Ross et al., 1992), (iii) long-distance transport of colluvial sediments (Nanson et al., 1988), or (iv) local accumulation of detrital material from the escarpment face and intervening scree slope (Young, 1988; Roberts, 1991).

In situ cosmogenic dating of the bedrock outcrops and profiles within the adjacent sand sheets may help to understand the processes of landscape evolution of sandstone landscapes in the Keep River region. More specifically, measurement of in situ cosmogenic nuclides in depth profiles can differentiate between extremes of autochthonous (in situ) and allochthonous formation of the sand sheets. Broad regional comparisons can be made between climatic parameters and landform assemblages (Derbyshire, 1976), and between long-term environmental and archaeological trends (Lourandos and David, 1998: 110). While the long-term archaeological trends for the Keep River region are yet to be determined, the inferred geomorphic environment can begin to provide a context for interpreting these trends.

The following application of in situ cosmogenic isotope dating in the Keep River region aims to:

- determine the denudation rate of the bedrock plateaux based on a direct model interpretation of bedrock cosmogenic $^{10}$Be and $^{26}$Al radionuclide concentrations,
- model the accumulation rate of the sand sheets from depth profiles of measured $^{10}$Be and $^{26}$Al radionuclide concentrations,
- compare and contrast estimated values of denudation and sedimentation with other semi-arid and monsoonal sandstone environments, and
- discuss the geoarchaeological implications of the above results.
4.3 Methodology

4.3.1 Field Sampling

Bedrock

A measure of maximum plateau denudation was determined for the bedrock outcrops which are considered to represent the source for the sand sheets in each of the site areas of Jinmium, Goorurarmum and Karlinga. Unfortunately, it was not feasible in this study to measure vertical erosion rates directly using \emph{in situ} cosmogenic dating. Rather the main use of the plateau denudation rates is in the modelling and burial dating of the adjacent sand sheets. A total of 15 bedrock samples were collected in July 2000 (recorded sequentially for each site as a COS-number). Approximately 1 kg of rock was removed from well-exposed, apparently stable, horizontal surfaces along the plateaux of the three main bedrock escarpments (Fig. 4.6). Samples were primarily chosen according to the principle of elevation such that the higher the surface the older it is. Different weathering surfaces (e.g. tafoni, case hardening) were sampled in order to give an estimate of the variation of denudation rates within a particular lithology and catchment area. The height of the plateau above the sand sheets was estimated from topographic surveys and the location of each sample site was determined from a Global Positioning System (GPS) (Fig. 4.5). Two bedrock samples from each site catchment (total of 6) were used for subsequent analyses. Details of each sample and location are given in Appendix A2.1.
Figure 4.5  Location map of cosmogenic sample sites, including bedrock samples (COS-G2, G4, J1, J2, K1 and K5) and sediment samples from the Jinmium sand sheet (JG1 and JG2).
Figure 4.6  Photograph of three of the bedrock samples (a) COS-J1, (b) COS-G2, and (c) COS-K1, collected for in situ cosmogenic dating. Samples were taken from horizontal surfaces with different forms of weathering.
Sand Sheets

In order to obtain a measure of the corresponding deposition history of the sand sheet sediments, two sediment cores were obtained from the Jinmium sand sheet. The Jinmium site was chosen as previous luminescence dating of a nearby auger site (Fullagar et al. 1996) indicated both the necessary antiquity (ca. 103 ± 14 ky BP) and depth (ca. 5 m) required for a $^{10}$Be and $^{26}$Al profile which may exhibit a shallow subsurface maximum and decay at depth. The first core (JG1) was limited to a depth of 3 m by a pisolithic sandstone base (refer A3.1). The second core (JG2) was augered to a depth of 6 m, which was the limit of the auger with no basal contact. Samples were taken at 30 cm intervals with a maximum depth error of ± 5 cm. Eight of the 10 samples were analysed from the 3 m core, and 10 of the possible 18 from the 6 m core, being a sufficient number to allow derivation of an accumulation rate. Permission to sample the plateaux and sand sheets was obtained from traditional custodians and no samples were taken from art surfaces. Details of each sample location are given in Appendix A2.1 and sediment descriptions given in Appendix A3.1.

4.3.2 Sampling Chemistry and AMS Measurement

Typically, only about a million atoms of $^{10}$Be and $^{26}$Al are produced in situ within a ten thousand year exposure period per gram of surface rock, hence the technique of Accelerator Mass Spectrometry (AMS) is needed to measure this signal. All bedrock and sediment samples were processed at ANSTO, following a three-stage process of sample preparation, AMS measurement, and data analysis (modelling) and interpretation. An overview of AMS techniques, laboratory practices, and discussion of exposure age applications is given by Tuniz et al. (1998).

Sample processing follows the ANTARES methods detailed in Child et al. (2000), which is based on a modification of Kohl and Nishiizumi (1992) and Brown et al. (1991). Raw sample masses ranging from 0.2 to 1 kg were used to obtain 50 – 100 grams of final pure quartz after chemical treatment. Following the sieved extraction of the 200 – 700 μm fraction from crushed bedrock or
sediment, samples were subjected to systematic selective leaching in dilute HF to obtain pure quartz as a closed system for in situ produced $^{10}$Be and $^{26}$Al. Selective leaching eliminates the potential atmospheric $^{10}$Be contamination and reduces the concentration of aluminium sourced from contaminant mineral fractions outside and within each quartz grain. Excess aluminium in etched quartz decreases the $^{26}$Al/$^{27}$Al ratio and hence sensitivity of $^{26}$Al as a chronometer. Constant $^{10}$Be concentrations were obtained after ~ 30 % etching, after which the residual quartz was completely dissolved in HF. The high purity of quartz meant there was no need for any heavy liquid separation of other contaminant minerals. The sample and HF mixture was then spiked with ~ 0.3 – 0.5 mg $^{9}$Be carrier, but no Al carrier as there were adequate concentrations in the clean quartz. ICP-AE results for Al range from 93 to 270 ppm, with typical errors of 2 – 4%. Be and Al were isolated, in the form of oxide, by solvent extraction as outlined in the flowchart by Child et al. (2000). The final pure powder samples were mixed with Nb powder and loaded into holders for AMS isotopic analyses.

Isotopic concentrations of $^{10}$Be and $^{26}$Al were measured using the ANTARES AMS spectrometer. Several measurements are made on any single target sample, and the weighted mean concentration is used to determine the denudation rate. An outline of the ANTARES AMS setup are given by Lawson et al. (2000), and details of AMS techniques and laboratory practices is given by Tuniz et al. (1998). The final AMS measurement errors depend predominantly on statistical errors of total $^{10}$Be and $^{26}$Al counts. Final quoted ratios are the weighted ratios of 4 – 6 independent repeat measurements. For $^{10}$Be/$^{9}$Be ratios these are typically < 5% for younger samples (~15 ka) and < 2% for older samples (> 100 ka) based on counting statistics, variability in measurement of standards, and uncertainty in blank corrections. For $^{26}$Al the errors are larger, around 5 – 10 %, due to lower $^{26}$Al count rates.
4.4 Results

4.4.1 Bedrock

The results of all AMS determinations for the six bedrock samples are given in Appendix A2.3 for \(^{10}\)Be and A2.4 for \(^{26}\)Al, along with the calculated isotopic concentrations, calculated site production rates, and denudation rates. A summary of the results is given in Table 4.1, and discussed further below.

Table 4.1 Summary of apparent exposure age and maximum denudation rate estimated from \(^{26}\)Al and \(^{10}\)Be isotope concentrations in six bedrock samples from the Keep River region. Refer 4.2.1 for definition of apparent exposure age (\(t_{\text{app}}\)). Results were calculated using a production rate of 5.1 at.g\(^{-1}\) for \(^{10}\)Be and 31.1 at.g\(^{-1}\) for \(^{26}\)Al at sea-level and high latitude (from Stone, 2000), and scaled to the relevant latitude and altitude using the scaling factors of Lal (1991).

Please see print copy for Table 4.1
The apparent exposure age, or the time required to remove ~ 500 mm of these geological surfaces, ranges from 75 to 210 ka for $^{26}$Al and 75 to 210 ka for $^{10}$Be (average ~ 125 ky BP). The average denudation rate is about 5.0 ± 0.7 mm.ka$^{-1}$ for $^{26}$Al and 4.8 ± 0.7 mm.ka$^{-1}$ for $^{10}$Be, and values range from 2.4 mm.ka$^{-1}$ (GUR4) to ~ 7.0 mm.ka$^{-1}$ (JIN2). The mean bedrock concentration of 4.9 ± 0.7 mm.ka$^{-1}$ is taken as the mean bedrock denudation rate over the past few Ma. With the exception of KAR5, and within the 10 % margin of error, the remaining five bedrock samples plot within the island between ‘constant exposure’ and ‘steady erosion’ (Fig. 4.7), indicating that they have undergone continual exposure with little or no burial or shielding of these bedrock samples. The position of KAR5 below the ‘constant exposure’ line may indicate some prior shielding.

Figure 4.7 $^{26}$Al/$^{10}$Be ratio diagram indicating position of bedrock samples in relation to the steady-state erosion island (see text for further explanation).
4.4.2 Sand Sheet Sediments

Raw results

The average $^{26}$Al and $^{10}$Be concentrations and $^{26}$Al / $^{10}$Be ratio per depth for cores JG1 and JG2 are given in Table 4.2 and shown graphically below (Fig. 4.8). Full results of all $^{10}$Be and $^{26}$Al analyses are detailed in Appendix A2.4 and A2.5 respectively.

Figure 4.8 Plot of $^{26}$Al / $^{10}$Be ratios for bedrock samples, and sediment samples from auger cores JG1 and JG2 on steady-state erosion island diagram. The position of the auger samples below the steady-state curve indicates that cores samples have not undergone cycles of burial and exposure.
Table 4.2  Concentrations and ratio of $^{10}$Be and $^{26}$Al radionuclides determined from successive analyses of target samples, taken from various depths in the two burial profiles (JG1 and JG2) near Jinmium. The high concentration of $^{10}$Be at 300 cm in the JG1 core (italicised) may represent a buried pediment surface (refer text for explanation).

<table>
<thead>
<tr>
<th>Profile Sample</th>
<th>Depth (cm)</th>
<th>$^{10}$Be (10^6 atoms/g)</th>
<th>$^{26}$Al (10^6 atoms/g)</th>
<th>$^{26}$Al / $^{10}$Be</th>
</tr>
</thead>
<tbody>
<tr>
<td>JG1 30</td>
<td>30</td>
<td>0.534</td>
<td>2.180</td>
<td>4.34</td>
</tr>
<tr>
<td>JG1 60</td>
<td>60</td>
<td>0.476</td>
<td>2.354</td>
<td>4.94</td>
</tr>
<tr>
<td>JG1 90</td>
<td>90</td>
<td>0.526</td>
<td>2.006</td>
<td>3.82</td>
</tr>
<tr>
<td>JG1 150</td>
<td>150</td>
<td>0.599</td>
<td>1.947</td>
<td>3.25</td>
</tr>
<tr>
<td>JG1 180</td>
<td>180</td>
<td>0.545</td>
<td>1.799</td>
<td>3.30</td>
</tr>
<tr>
<td>JG1 210</td>
<td>210</td>
<td>0.539</td>
<td>2.154</td>
<td>3.99</td>
</tr>
<tr>
<td>JG1 270</td>
<td>270</td>
<td>0.591</td>
<td>1.699</td>
<td>2.88</td>
</tr>
<tr>
<td>JG1 300</td>
<td>300</td>
<td>0.705</td>
<td>1.644</td>
<td>2.33</td>
</tr>
<tr>
<td>JG2 30</td>
<td>30</td>
<td>0.584</td>
<td>2.775</td>
<td>4.75</td>
</tr>
<tr>
<td>JG2 60</td>
<td>60</td>
<td>0.629</td>
<td>2.827</td>
<td>4.49</td>
</tr>
<tr>
<td>JG2 90</td>
<td>90</td>
<td>0.634</td>
<td>2.958</td>
<td>4.67</td>
</tr>
<tr>
<td>JG2 150</td>
<td>150</td>
<td>0.632</td>
<td>2.829</td>
<td>4.47</td>
</tr>
<tr>
<td>JG2 210</td>
<td>210</td>
<td>0.685</td>
<td>2.579</td>
<td>3.77</td>
</tr>
<tr>
<td>JG2 300</td>
<td>300</td>
<td>0.599</td>
<td>2.508</td>
<td>4.19</td>
</tr>
<tr>
<td>JG2 360</td>
<td>360</td>
<td>0.589</td>
<td>2.456</td>
<td>4.17</td>
</tr>
<tr>
<td>JG2 450</td>
<td>450</td>
<td>0.627</td>
<td>2.224</td>
<td>3.55</td>
</tr>
<tr>
<td>JG2 540</td>
<td>540</td>
<td>0.584</td>
<td>1.716</td>
<td>3.46</td>
</tr>
<tr>
<td>JG2 600</td>
<td>600</td>
<td>0.526</td>
<td>1.086</td>
<td>3.17</td>
</tr>
</tbody>
</table>

In the burial profiles of JG1 and JG2, the $^{26}$Al/$^{10}$Be ratio decrease steadily with depth (Table 4.2). The mean concentration of $^{26}$Al and $^{10}$Be in the shallow surface (< 100 cm) is ~ $2.5 \times 10^6$ at.g$^{-1}$ and ~ $0.56 \times 10^6$ at.g$^{-1}$ respectively. These values are similar to the mean bedrock concentrations (Table 4.1), supporting the assertion that the local bedrock is the main source of sediment to the sand.
sheets. The consequent mean ratio of $^{26}$Al/$^{10}$Be in the shallow surface is $\sim$ 4.5 and is less than the mean bedrock $^{26}$Al/$^{10}$Be ratio. The shorter JG1 profile shows an irregular increase of $^{26}$Al at 210 cm (Table 4.2). The base of JG1 comprised pisolithic sandy sediments, which differed to the unconsolidated red sands above. The comparatively elevated concentration of $^{10}$Be at the base (300 cm) of the JG1 profile (Table 4.2), may represent inheritance preserved in a previously exposed surface or pediment, with comparable cosmogenic isotope concentrations to the presently exposed bedrock. For the longer JG2 profile, the concentrations of $^{10}$Be/$^{26}$Al ratio shows a gradual decrease with depth.

Estimating $E_0$ and $S$

For the two auger cores JG1 and JG2, sedimentation rate is determined by using the weighted mean of measured $^{10}$Be and $^{26}$Al concentrations as the initial value for deposition and modelled with depth profiles using eq. (7) with a range of approximated accumulation rates ($S$) (refer 4.2.3). Model results were calculated using a value for P(0) of 3.79 at.g yr$^{-1}$ and 19.90 at.g yr$^{-1}$ for $^{10}$Be and $^{26}$Al respectively (15.43°S, 10 m altitude), a density of 1.66 g.cm$^{-3}$, an attenuation length of 150 g.cm$^{-2}$, and a decay constant of $4.59 \times 10^{-7}$ y$^{-1}$ for $^{10}$Be and $9.89 \times 10^{-7}$ y$^{-1}$ for $^{26}$Al. In order to match measured concentration with modelled profiles for $^{10}$Be and $^{26}$Al, the best correspondence was obtained by using the average concentrations of $0.41 \times 10^6$ at.g$^{-1}$ and $2.42 \times 10^6$ at.g$^{-1}$ respectively, rather than the minimum or maximum bedrock concentration (from 4.4.1).

Fig. 4.9 shows the profiles for sedimentation rates of 5, 10, and 20 mm.ka$^{-1}$ for $^{10}$Be. With the exception of the sample at 60 cm (disturbance) and 300 cm (which may represent buried bedrock), the measured $^{10}$Be concentrations in JG1 correspond to a sedimentation rate between 8 and 15 mm.ka$^{-1}$. For the deeper core JG2, the measured profile indicates a sedimentation rate between 5 and 10 mm.ka$^{-1}$. In summary, the measured concentrations for both $^{10}$Be profiles are bracketed by the model curves with accumulation rates of 5 – 20 mm.ka$^{-1}$. Any lower sedimentation rates ($S$)
would predict a larger maximum and sharper fall off at depth, while any higher $S$ would lie well below all the data points, effectively invariant with depth.

For $^{26}$Al the correspondence between the measured concentrations and the model curves for JG2 is acceptable but is poorer for JG1. The measured concentrations in the JG1 core fall below any of the model curves and are difficult to explain. The profile for JG2 corresponds to a sedimentation rate between 10 mm.ka$^{-1}$ and 20 mm.ka$^{-1}$, which is larger than that determined from $^{10}$Be. Attempts to force a better fit to the profiles of JG1 and JG2 (using an equivalent $S$ of 5 – 10 mm.ka$^{-1}$ for $^{10}$Be), using a lower initial $^{26}$Al concentration ($C_0$) are unsuccessful. For $S > 20$ mm.ka$^{-1}$, the curves are flat and require a much higher initial $^{26}$Al concentration ($C_0$). Thus using $^{26}$Al the average rate of sedimentation is also estimated to be 10 – 20 mm.ka$^{-1}$.

It is also necessary to consider the possible effects of bioturbation on the burial profiles. Using the theory of homogenisation (Braucher et al., 2000) the extent of bioturbation is estimated to be less than 100 cm. This is supported by comparable mean concentrations of $^{26}$Al and $^{10}$Be over this depth and the source bedrock, and by the relatively poor fit of the model profiles over this depth. Beyond this depth the measured concentrations, although scattered, do show some decay with depth. However, the cosmogenic data show two features which indicate that vertical mixing may be significant: (i) more scatter than predicted from theory (compare Figs. 4.3 and 4.9) and (ii) depressed $^{26}$Al/$^{10}$Be ratios (4.3 – 4.8) in near-surface samples given their inferred sedimentation rate ($\sim 10$ mm.ka$^{-1}$).

In summary, although the measured concentrations provide a relatively flat curve, they typically show weak subsurface maxima and tail off at depths below 300 cm. The cosmogenic data are sensitive to prior erosion rate ($\sim 5$ mm.ka$^{-1}$), which provides a representative accumulation rate of 10 – 20 mm.ka$^{-1}$ for the sand sheets. Maximum errors have been integrated into the cosmogenic data, but there remains some uncertainty regarding the sedimentation rate. Ironically, burial dating becomes less precise with increasing erosion rates.
Figure 4.9 Measured radionuclide concentrations for (a) $^{10}\text{Be}$ and (b) $^{26}\text{Al}$ for the two cores JG1 and JG2, plotted against model profiles for sedimentation rates of 5, 10, 20 and 50 mm.ka$^{-1}$, using $C_0$ calculated from the average bedrock erosion rate ($E_{\text{ave}}$). Refer text for explanation.
4.5 Discussion

The use of cosmogenic radionuclides to determine sedimentary processes in buried sediments follows the concept developed by Brown et al. (1994) on buried laterite. The work also complements in situ cosmogenic isotopic studies already being undertaken by Fink et al. (2000), and by other researchers in Arnhem Land (Nott and Roberts, 1996), and Flinders Ranges (Heimsath et al., 2000). The following discusses the estimates of bedrock denudation and sand sheet accumulation in the Keep River region, in the context of landscape weathering and sand sheet evolution. A hypothetical model of sand sheet evolution dominated by long-term deep weathering is proposed, with the aim of providing a context for interpreting the archaeological trends for the Keep River region.

4.5.1 Landscape Weathering over the Past 200 ka

As outlined in Chapter Three, the climatic environment since the early Pleistocene has changed from being cool and wet, to one that was increasingly arid up to the time of the Last Glacial Maximum (LGM), to the semi-arid monsoonal climate of today. Changing environmental conditions have been critical to the formation and differential erosion of sandstone assemblages, but the manifest erosional variation within sandstone assemblages is generally greater than that between similar rock-types under different climatic regimes (Young and Young, 1992). In the Keep River region, the measured ratios of $^{10}$Be and $^{26}$Al in the six bedrock samples fall near the expected steady-state values, indicating continual exposure of unshielded bedrock resulting in erosional equilibrium conditions. As argued for Arnhem Land (Roberts, 1991), the weathering rate of the escarpment sandstone in the Keep River region is apt to be slow and reasonably constant under any climatic regime.

Model minimum exposure ages of the bedrock plateaux in the Keep River average around 125 ka, and in the context of exposure age modelling, corresponds to the time taken to denude
approximately 500 mm of bedrock (refer 4.2.1). The period of denudation extends back to the last
interglacial period of the early Pleistocene. The measured denudation rate over this time period
averages about 5 mm.ka\(^{-1}\), regardless of the presence of case-hardened surfaces or otherwise,
indicating that there is little temporal or spatial variation in denudation rate at this scale of analysis.
It should also be remembered that denudation rates based on cosmogenic radionuclides average
several hundred thousand years of exposure and erosion and are buffered against short term changes
in denudation rates (Brown et al., 1995). Cosmogenic nuclide concentrations thus reflect pre-
anthropogenic erosional conditions (Brown et al., 1998).

Recent work by Bierman and Caffee (2002) indicate that denudation rates are slightly higher (i.e.
radionuclide concentrations are lower) in northern Australia than in other parts of Australia,
inferring that surface stability is inversely related to mean annual precipitation. The denudation
rates calculated from over 60 exposed bedrock surfaces ranged from 0.3 mm.ka\(^{-1}\) from granite
inselbergs to 5.7 mm.ka\(^{-1}\) on sandstones (Bierman and Caffee, 2002). The estimates of plateau
denudation for the Keep River region are similar to those estimated from \textit{in situ} cosmogenic dating
in the Arnhem Land plateau (Roberts, 1991; Nott and Roberts, 1996), Victoria Plateau (Fink et al.,
2000), Flinders, Macdonnell and Mt Isa Ranges (Heimsath et al., 2000), Eyre Peninsula (Bierman
and Turner, 1995; Bierman and Caffee, 2002), and also in the climatically similar environment of
South Africa (Flemming et al., 1999). These cosmogenic-based erosion rates also correspond to
estimates for semi-arid climates obtained from a review of numerous sources by Young (1983),
which range from about 10 mm.ka\(^{-1}\) under humid temperate climates, 1 mm.ka\(^{-1}\) for semi-arid, and
0.01 mm.ka\(^{-1}\) for arid climates. Thus the estimated denudation rates for Keep River are regionally
consistent.

Erosion of landscape locally and over shorter periods (decades to millenia) may be more erratic and
based on stochastic events rather than through a slow process of weathering of the bedrock surface.
After the 2000 wet season, which had over 250 mm of rainfall, evidence of a landslide or rock fall
was visible on the Goorurarmum escarpment (Fig. 4.10). The effect of such a brief episodic event
can outweigh the mass transport resulting from continuous spalling of the escarpments and rock-fall debris, but generally only where such events are both numerous and the evidence is obvious (Saunders and Young, 1983). The block-wise denudation of landscapes is presently being considered by Chappell et al. (2001) in the sandstone terrain of the Flinders Ranges, and give apparent erosion rates that are slightly greater (mean $E_{\text{app}} = 166 \text{ mm.ka}^{-1}$) than those estimated from long-term uniform erosion (mean $E_{\text{app}} = 143 \text{ mm.ka}^{-1}$). Thus erosion rates in some parts of the Keep River region may easily be greater than $10 – 20 \text{ mm.ka}^{-1}$.

Figure 4.10 Landslip off the Goorurarmum escarpment that occurred sometime after the 2000 wet season, looking towards Sandy Creek Gorge.

Many months after the end of the 2000 wet season there was evident seepage from the escarpment faces (Fig. 4.11), indicating the likely saturation and chemical disintegration of the bedrock through loss of silicate clay minerals (Young and Young, 1992). Removal of intergranular cement can reduce the mechanical strength, while case hardening through re-deposition of silica or iron oxides near the surface may increase strength (Young and Young, 1992).
Figure 4.11  Seepage of water through several metres of bedrock was evident several months after the end of the 2000 wet season, indicating the likely gradual saturation and disintegration of the bedrock through chemical weathering.

This differential cycle of disintegration and stabilization is evident from the deeply etched stratigraphic units, exfoliating slabs, tafoni, and case-hardened and lateritic surfaces throughout the Keep River region. Although permanence was not the purpose of most Aboriginal art (Mulvaney and Kamminga, 1999), it is possible that Aboriginal people deliberately chose to place their art on more stable surfaces. These surfaces would also have been subject to differential erosion and preservation. Consequently, the erosion rate of any local rock-art surface may not be represented by regional estimates of vertical or horizontal denudation, nor should the age of the art be reckoned from these regional estimates.
4.5.2 Sand Sheet Evolution

The cosmogenic depth profiles for $^{10}$Be and $^{26}$Al (Fig. 4.9) are clearly more typical of (allochthonous) accumulating sediments rather than a burial profile developed from in situ (autochthonous) weathering. Furthermore, the correlation between the cosmogenic isotopic concentration of the bedrock and surface sediment samples support the assumption that the sand sheet sediments are locally derived. Thus the results from in situ cosmogenic dating support the formation of the sand sheets from local accumulation of detrital material from the escarpment face and intervening scree slope in line with Young (1988) and Roberts (1991).

The higher basal concentration of $^{10}$Be at the base of the JG1 profile is interpreted to represent a previously exposed palaeosurface upon which later sediments accumulated. Interpolating the estimated accumulation rate of 10 – 20 mm.ka$^{-1}$ of the sand sheet sediments to a minimum age of the inferred palaeosurface at 300 cm in JG1, gives an estimated age of 300 – 150 ka respectively. This is not unreasonable given previous TL ages for these sand sheet sediments in the order of 100 ka (Fullagar et al., 1996).

A hypothetical reconstruction of long-term sand sheet evolution in the Keep River region is illustrated in Fig. 4.12, with the model illustrating the initial accretion of sand sheets over a previously exposed palaeosurface. The accumulation of sand over this palaeosurface is likely to be discontiguous with earlier (older) sediments filling and being preferentially preserved in depositional lows compared to later (younger) sediments. Maxwell and Haynes (2001) describe a comparable formation history for the evolution of the Selima sand sheet, in Egypt. In both cases, older archaeological sequences are more likely to be represented in these topographic lows.

Subsequent accumulation of the sand sheets may occur through gradual and continuous processes, or may occur through major catastrophic events of accretion and partial denudation, akin to the ‘episodic disequilibrium’ described by Nanson (1986). In parts of Australia, periodic expansion of
sand dunes and sand sheets occurs through aeolian activity during acutely arid glacial stages (Bowler, 1976; Pye and Gardner, 1981; Ross et al., 1992; Wasson, 1983, 1986; Wende et al., 1997), whilst in other parts, expansion has occurred as a result of alluvial activity during interglacials (Nanson et al., 1988). At Cabbage Creek, approximately 150 km south of Kununurra, evidence of enhanced alluvial activity is indicated from dated overbank deposits at 37 ka, and between 12 and 6 – 5 ka (Wende et al., 1997). Thus it is quite possible that some expansion of sand sheets also occurred in the Keep River region in association with these wetter interglacial periods.

Contrasting episodes of sand sheet denudation are indicated from geomorphological studies on the Arnhem Land plateau and Magela Creek catchment (Roberts, 1991; Nanson et al., 1993) and in the nearby Koolpin Gorge (Walsh, 1993). All these studies indicate major episodes of alluvial stripping occurred prior to Stage 5, with later events of partial stripping since 100 ky BP, particularly during Stage 3. Geomorphological studies in the Arnhem Land region indicate that these periods of enhanced denudation most likely resulted from dramatic changes in sea-level and/or more frequent rainfall events and floods (Nott and Price, 1994, 1999; Nott and Roberts, 1996). Situated much closer to the coast than Arnhem Land, the Keep River region would have been even more vulnerable to such erosional events, resulting in more frequent and/or extensive episodes of sand removal over the past 100 ka and possibly throughout the entire Quaternary.

During periods of relative stability or where accumulation is more gradual and continuous, soil-forming processes may allow distinct horizons to form. In the White Paintings rock shelter, in the northwest Kalahari Desert, soil A-horizon development not only delimited particular sand units but also correlated with significant peaks in artefact materials (Robbins et al., 2000). However, in northern Australia quartz-rich sands in the form of dunes and sand plains have often been found unfavourable for the formation and preservation of any internal sedimentary structures (Mulvaney and Kamminga, 1999; Newsome, 2000) or palaeosol horizons (Bowler et al., 2001). Field observations in the Keep River region indicate that this is also true of the quartz-rich sand sheets.
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Figure 4.12 Hypothetical reconstruction of sand sheet evolution over timescales of several millennia. Initial accretion of sand sheets over an exposed pediment occurs through deep weathering and erosion of the escarpment. Subsequent vertical accretion may occur through (a) gradual accumulation of sand, (b) episodic events of erosion and sand deposition, or (c) autochthonous weathering. Human occupation may occur during any period of relative landscape stability or instability. Refer text for further explanation.
Thus over time, net deposition may occur with or without obvious major facies and/or stratigraphic boundaries, which may only be determined from the chronostratigraphy or micromorphology. Unfortunately, the chronostratigraphic resolution from *in situ* cosmogenic dating is not favourable, in these relatively young and shallow sequences because the estimated sedimentation rate (S) is a mean for the entire period of sand sheet accumulation. Overall, the results of *in situ* cosmogenic dating indicate that the evolution of the sand-sheets in the Keep River region over the past several hundred thousand years has been gradual (10 - 20 mm.ka$^{-1}$), involving local derivation of weathered materials and colluvial reworking and bioturbation.

### 4.6 Validating Denudation and Accumulation Rates from Cosmogenic Dating

The estimated rate of plateaux denudation from exposure dating (~ 5 mm.ka$^{-1}$) is less than the estimated rate of sediment accumulation from burial dating (10 - 20 mm.ka$^{-1}$) in the adjacent sand sheets. However, the two measures are not directly comparable. Studies in Arnhem Land indicate that only 5% of the volume of the sand sheets comprises sand denuded from the plateau surface and the remaining 95% derives from the escarpment face (Roberts, 1991). From $^{10}$Be and $^{26}$Al cosmogenic dating, calculated denudation rates for the Arnhem Land plateaux of 4 mm.ka$^{-1}$ (Nott and Roberts, 1996) contrast with scarp retreat rates estimated from volumetric calculations of 20 to 200 mm.ky$^{-1}$ (Roberts, 1991). On the Drakensberg escarpment, SE South Africa, Fleming et al. (1999) used cosmogenic $^{36}$Cl abundances to measure plateaux denudation rates of 5 mm.ka$^{-1}$ and scarp retreat rates of 50 – 95 mm.ka$^{-1}$. Weathering of these basalt escarpments also occurred through regular spalling of thin sheets or individual grains and/or the periodic loss of half-metre blocks. Comparable estimates of plateau denudation of 5 mm.ka$^{-1}$ (Heimsath et al., 2000) and block-wise retreat of 166 mm.ka$^{-1}$ (Chappell et al., 2001) are obtained from sandstone landforms in the Flinders Ranges. Thus assuming similar ratios also exist in the Keep River region, it is possible to infer escarpment erosion rates in the order of 50 - 100 mm.ka$^{-1}$. 
Correspondingly, sediment accumulation ($S_{\text{area}}$) can be estimated according to the relative areas of supply ($A_{\text{bed}}$) and deposition ($A_{\text{ss}}$), assuming that these bedrock outcrops ($E_{\text{bed}}$) are the only source for the adjacent sand sheets:

$$S_{\text{area}} = \frac{\text{Erosion rate} \times \text{Area of bedrock}}{\text{Area of sand sheet}}$$

The area of contributing escarpment in each of the catchment areas of Karlinga and Goorurarmum is estimated from topographic maps (refer Fig. 4.5) to be in the order of 2 km$^2$. Converting the bedrock to its granular equivalent by adding 20% for inter-granular porosity (Roberts, 1991) increases the effective area of the bedrock source to 2.4 km$^2$. The adjacent sand sheets in each local catchment area can be approximated a length and breadth of 4 km (refer Fig 3.5), giving an areal extent ($A_{\text{ss}}$) of 16 km$^2$. Assuming an erosion rate of 100 mm.ka$^{-1}$ gives an estimated sedimentation rate ($S_{\text{area}}$) of around 14 mm.ka$^{-1}$, which is comparable to sedimentation rates estimated from cosmogenic burial dating ($S_{\text{cosmo}}$). The equivalence of the sedimentation rate estimated by cosmogenic dating ($S_{\text{cosmo}}$) and from geometric determinations ($S_{\text{area}}$) supports the assumption that the local bedrock outcrops are the main or only source of material to the adjacent sand sheets.

These assumptions are given further credence in the following chapters on luminescence dating and sediment characterisation.
4.7 Geoarchaeological Implications

Modelling, measuring and comparing regional archaeological trends have presented problems for the archaeologist (Lourandos and David, 1998). Modelling, measuring and comparing regional geological processes are perhaps first steps towards defining the dynamics of the human landscape. As the physical processes shaping the land can be seen to unfold over hundreds and thousands of years, so can the social processes which are evident in the archaeological data (Gosden and Head, 1994: 113). Geomorphodynamics in dry savanna environments can often have both anthropogenic and climatic causes (Heinrich and Moldenhauer, 2002).

In a recent program on the Kimberley rock art, an allusion was made to the weathering rate of the sandstone bedrock. It was stated that:

“In the Kimberley, and in Arnhem Land, the rock is so hard, you can actually see paintings that are 40 to 50 thousand years old still on the surface.”

(M. Morwood on ‘Australian Story’ ABC, 14 October 2002).

The statement implies that the erosion rate of the sandstone bedrock is less than the thickness of the paintings, or less than $1 - 2 \text{ mm per } 50 \text{ ka or } 0.05 \text{ mm.ka}^{-1}$. Such notional statements highlight the need for more empirical quantification of denudation and erosion rates of sandstone surfaces throughout the Kimberley and Arnhem Land. In the Keep River region, as in other parts of northern Australia (refer 4.5.1), in situ cosmogenic isotope dating provides an estimation of plateau denudation that is $1 - 2$ orders of magnitude greater than the above implied erosion rate of the escarpment faces upon which the rock art occurs. However, it should be remembered that in situ cosmogenic dating averages processes occurring over million year timescales and not the hundred or thousand year timescales over which local weathering of rock art occurs. In other words, higher resolution dating techniques may provide better estimates of weathering of local rock art surfaces (see also Bednarik, 2002; Watchman and Twidale, 2002).
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The results from in situ cosmogenic isotope dating indicate that the physical processes shaping the sandstone landscapes in the Keep River region date well beyond 100 ky BP, and compares with the archaeological history of northern Australia with occupation dates around 60 ky BP (Roberts et al., 1994b; Turney et al., 2001). Although burial dating necessarily assumes gradual accumulation of the sand sheets, a hypothetical model proposes that sand sheet formation occurs episodically, with sediment and associated archaeological deposits preserved accordingly. In both the physical and human landscape are shorter term dynamics. The shorter term landscape dynamics of the Keep River region are the subject of the following chapters on luminescence dating and sediment characterisation.

4.8 Conclusions

Using in situ cosmogenic nuclide dating, the estimated rate of plateau denudation is 5 mm.ka\(^{-1}\) over the past 500 ka, which corresponds well to regional estimates. However, greater resolution is required to accommodate recent catastrophic events (e.g. rock falls) or estimate erosion at specific site locations, particularly for surfaces where rock art observed.

The bedrock denudation rates are used to model depth profiles of \(^{10}\text{Be}\) and \(^{26}\text{Al}\) concentrations through the sand sheets. The results support the formation of sand sheets through local accumulation of detrital material from the escarpment face and intervening scree slope rather than from in situ conversion of the underlying saprolite or long-distance transport of aeolian sediments. The local sourcing of material to the sand sheets is indicated from the correspondence between the isotopic concentration in the bedrock and surface sediments. Burial dating indicates that over the past several hundred thousand years, accumulation of sand in the sand-sheets has been greater than erosion with a net sedimentation rate of 10 - 20 mm.ka\(^{-1}\). However, the resolution from these sediments cannot discern intermediate changes in accumulation rate or episodes of erosion over shorter timescales.