Deriving basin-wide denudation rates from cosmogenic radionuclides, San Bernardino Mountains, California

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This thesis has been composed entirely by myself and is my own work. This work has not been submitted for any other degree or professional qualification.

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The structure of this thesis is such that chapters 4, 6 and 7 should be viewed as initial drafts of papers intended for future publication. These chapters can, therefore, be read independently of the others. However, this requires that there be some repetition of background material and figures in these chapters.
Abstract

As increasing emphasis is placed upon the role surface processes play in regulating tectonic behaviour, the need for accurate measurements of denudation rate has become paramount. The quantity and quality of denudation rate studies has grown with the advent of cosmogenic radionuclide techniques, capable of recording denudation rates over timescales of $10^2$ to $10^6$ years. This study seeks to utilise cosmogenic $^{10}$Be concentrations measured in alluvial sediments in order to further develop this method and to investigate rates of basin-wide denudation in the San Bernardino Mountains, an active orogen associated with the San Andreas Fault system. The theory which underpins measurements of basin-wide denudation rates with cosmogenic radionuclide analysis is evaluated in light of recent understanding of production mechanisms. Field testing of the assumptions required by the basin-wide denudation rate model highlights the importance of sampling thoroughly mixed sediments. Denudation rates ranging over three orders of magnitude are measured by applying the cosmogenic radionuclide technique in thirty-seven basins throughout the San Bernardino Mountains. Results show a relationship between
denudation rate and slope which provides quantification of the threshold slope angle in high relief granitic environments and suggests tectonic activity is the first order control of denudation rates in these mountains. Mean annual precipitation is shown to exert no significant influence over the rates measured in the San Bernardino Mountains. Questions concerning denudation rates recorded over differing time-spans are addressed using the cosmogenic technique, (U-Th)/He thermochronometry, incision into dated horizons and modern day sediment flux data. This comparison reveals that a decrease in rates with distance from the San Andreas Fault has been consistent throughout the lifespan of the San Bernardino Mountains and provides further evidence that a tectonic mechanism is driving denudation in this region. The relevance of both spatial and temporal scale in geomorphic studies is considered in light of these results, highlighting the need for a greater appreciation of their role in the interpretation of basin-wide denudation rates.
1. Introduction

1.1 Denudation in the earth sciences:

The rates at which geomorphic processes operate at the earth’s surface are fundamental to an understanding of landscape evolution. In particular, recent focus on the role of denudation in orogenesis has been driven partly by the recognition that surface process rates have the potential to modulate the operation of global climate and tectonics (Merritts and Ellis, 1994; Pinter and Brandon, 1997; Summerfield, 2000). The interdisciplinary approach required to tackle such issues has helped define a tectonic-geomorphological framework in which the earth’s surface is viewed as an interface between internal, geological and geophysical processes and external, atmospheric and biological systems. As such, the rate at which landscapes are denuded acts to dampen or amplify tectonic and climatic change through feedbacks, hence influencing the form of mountainous landscapes not only through shaping the surface but by driving the agents that are generating or destroying relief. However, several key geomorphological questions have yet to be resolved before the evolution of mountain ranges can be integrated with notions of global climatic
change and tectonic theory. These include questions as to the factors driving
denudation rates (Milliman and Syvitski, 1992; Molnar, 2003); the mechanics of
surface processes operating over long time scales (Hovius et al., 1997; Stock and
Montgomery, 1999) and the rates and scales at which landscapes respond to change
(Vandeberghe, 2003; Burbank and Pinter, 1999). While several studies have
attempted to interpret these problems in older mountains ranges (Burbank et al.,
1996; 2003; Montgomery, 1994; Adams, 1980; Riebe et al., 2000) there has been
relatively little work focused on the very early stages of orogenesis, although it is
these early stages which can define the future form of the range (Hovius et al.,
1998). In order to answer these questions and hence develop, test and compare
theories of orogenesis within a tectonic-geomorphological arena, accurate estimates
of denudation from active, recently uplifted mountains are vital.

Measuring denudation rates over the $10^2$-$10^6$ year time scales that are pertinent to the
macroscale evolution of landscapes has proved to be difficult. Previously,
denudation rates derived over annual or decadal spans were extrapolated to the
longer-term in order to inform landscape evolution studies (Schumm, 1963; Anhert,
1970). This methodology is hazardous because it requires an assumption that
denudation rates at one time scale approximate those at another (Pinet and Souriau,
1988). Alternatively, if the right circumstances are present, it has been possible to
use dated horizons and sediment volumes in order to estimate denudation rates. This
allows longer-term issues of landform development to be addressed using
denudation rates measured over the appropriate period of time (Summerfield and
Hulton, 1994). However, the use of this method has been restricted by the
requirement for specific conditions. The emergence of cosmogenic radionuclide analysis in the mid-1980s, as a method whereby rates of bedrock denudation can be derived over $10^2$-$10^6$ year periods, presented a widely applicable geomorphological tool able to tackle questions of landscape evolution over the appropriate temporal scales (Lal, 1991). The further work of Bierman and Steig (1996), Granger et al. (1996) and Brown et al. (1995) developed an application of cosmogenic radionuclide analysis that allows the derivation of basin-wide denudation rates. This method is able to derive the spatially and temporally integrated denudation rates of entire drainage basins using the cosmogenic radionuclide concentrations of alluvial sediment. Therefore, cosmogenic radionuclide analysis presents a technique able to connect short-term sediment yield estimates with long-term thermochronometric measures of denudation and, as such, it provides the ability to compile denudation rate chronologies for landscapes over annual to geological time scales.

1.2 Denudation of the San Bernardino Mountains:

Many of the recent insights into the processes of orogenesis, and the role that denudation plays, have been based on long-term denudation rates derived from thermochronometric techniques which require the removal of several kilometres of overlying bedrock (Zeitler et al., 1987). While more developed mountain ranges will have denuded this depth of overburden many nascent mountains, in the early stages of orogenesis, will not. In such cases, the applicability of thermochronometric techniques to interpret the initial stages of mountain building will be restricted to those areas experiencing extremely rapid denudation. Cosmogenic radionuclide analysis presents an alternative method by which to
measure the rates of denudation over appropriate time scales during early stages of orogenesis.

In this study, the tectonic geomorphology of a young mountain range is addressed utilising cosmogenic radionuclide analysis to derive basin-wide rates of denudation. The San Bernardino Mountains in southern California provide the ideal location for this work. They are active, showing evidence of late-Pliocene uplift, and estimates of denudation rates derived over the lifespan of the mountains act as a baseline for comparison with the rates derived using the cosmogenic technique. There are steep climatic, structural and morphological gradients across the range but a relatively uniform lithology, allowing investigation of the effect that climatic, tectonic and geomorphic variables may have on rates of denudation. The San Bernardino Mountains lie on the extensively studied San Andreas Fault Zone, providing a well documented geological synthesis within which to perform this study.

1.3 Thesis layout:

The feasibility of using cosmogenic radionuclide analysis to derive denudation rates has been tested through its application. However, the application of cosmogenic radionuclide analysis in rapidly denuding, tectonically active regions has been limited (Vance et al., 2003). Because of this there are several unresolved issues relating to the use of the technique in such environments where rapid mass movement is prevalent. To address this shortcoming, Chapters 2, 3 and 4 of this thesis are devoted to the use of the technique and the assumptions it requires. Chapter 2 provides a general history of the development of cosmogenic radionuclide
analysis. Particular attention is paid to the basin-wide approach, whereby spatially-
averaged denudation rates are derived from sediment samples. As this approach is
the focus of subsequent chapters a review of its previous applications is provided. In
Chapter 3 the methodology behind cosmogenic radionuclide analysis is introduced.
The theory and assumptions of the basin-wide approach receives explicit attention.
This chapter considers the variables used in denudation rate calculations and the
scaling functions applied to them. Finally, Chapter 3 explains the laboratory
procedures needed to process sediment samples for analysis by accelerator mass
spectrometry. Chapter 4 presents the field-testing of some of the assumptions
described in Chapter 3. The focus of this testing is the assumptions of sediment
mixing made when using the basin-wide approach and the influence that mixing
may have on measurements of denudation rates. Also considered is the potential
bias in cosmogenic radionuclide concentrations with different sediment grain sizes
and the influence that denudational processes have when applying basin-wide
cosmogenic radionuclide analysis in rapidly denuding environments.

After the explanation and testing of cosmogenic radionuclide analysis the discussion
focuses on the San Bernardino Mountains and the interpretation of denudation rates
measured there using cosmogenic $^{10}$Be. Chapter 5 introduces the mountains as the
product of young orogenesis on the San Andreas Fault Zone and the origins,
structure and timing of uplift are discussed, together with the geomorphology and
palaeoclimate of the range. Previous work relating to the tectonic geomorphology of
the mountains is also reviewed. Chapter 6 investigates the controls of denudation
rates in the San Bernardino Mountains by addressing variables such as climate,
tectonic uplift and lithology and considering the influence of each. This analysis is undertaken with respect to different denudational regimes by dividing the sample sites into areas where landsliding is occurring, from those where it is not. Results quantify a relationship between slope gradient and denudation that decouples as slope gradient increases. The point at which this relationship decouples coincides with a change in the denudational processes operating within basins, from fluvial incision to landsliding. Furthermore, a lack of climatic control on denudation rates is recorded for the environments of the San Bernardino Mountains. Chapter 7 utilises previous studies of denudation rates in the San Bernardino Mountains over periods both longer and shorter than the averaging time of cosmogenic radionuclide analysis to provide an insight into denudation rates across a range of time scales. A general consistency between rates recorded over time scales which vary by five orders of magnitude is found. Chapter 8 provides a general discussion of the results from Chapters 3, 4, 6 and 7, expanding upon and developing the arguments made earlier. The implications from the field-testing of basin-wide cosmogenic radionuclide analysis are considered in relation to the rates measured in the San Bernardino Mountains. The controls of denudation are discussed in relation to thresholds within the geomorphic system and the importance of spatial and temporal scale in denudation rate studies is highlighted. The implications of the work presented here for the evolution of the San Bernardino Mountains and theories of orogenesis of young mountains is discussed. Finally, future areas of research employing the cosmogenic technique, and its applicability to studies of mountain building are suggested.
1.4 Terminology:

1.4.1 Erosion, denudation and exhumation:
Throughout this study the term denudation is used to refer to the wearing away of the earth’s surface through the operation of all processes of weathering and mass movement. In other studies the term erosion has been used to describe this. However, erosion specifically refers only to the mechanical destruction of the land, excluding transport and weathering, and accordingly the term will be reserved for this purpose. Cosmogenic radionuclide analysis measures the rate at which mass is removed from the earth’s surface, incorporating both the chemical and physical removal of material, and as such it measures overall denudation. Furthermore, the term exhumation is often employed in thermochronometric studies to indicate the process by which a point in the rock column will approach the earth’s surface. However, the process by which a point in the rock column will approach the surface is the denudation of overlying bedrock, whereas the term exhumation implies the uncovering or exposure of a pre-existing surface or feature (Summerfield and Brown, 1998). Accordingly, the term denudation will also be used rather than exhumation and if derived by thermochronometric techniques it will be referred to as long-term denudation.

1.4.2 Concepts of uplift:
The term uplift has become confused in earth science literature, due partly to its usage by thermochronometric studies when describing the upward motion of rock relative to the earth’s surface (bedrock uplift) (Summerfield and Brown, 1998). Here, the formation of the San Bernardino Mountains are described in terms of: 1.
the relative motion of crustal blocks in relation to their surroundings (relative uplift). This does not reveal surface uplift with respect to the geoid but documents the amount of displacement in a vertical frame of reference. Whether this displacement represents uplift or subsidence across a fault is not necessarily known, however, both scenarios will create relief and contribute to the formation of mountainous topography. 2. The elevations of palaeosnowlines are used as evidence of vertical crustal movement. In this case the motion is relative to sea-level, not the surrounding landscape (surface uplift). 3. The thermal history of rocks now at the surface, obtained using thermochronometric techniques, can reveal the rates at which the surface has denuded with respect to the thermal gradient of the crust (bedrock uplift). As such, changes in denudation rates from thermochronometric studies may be used to infer, indirectly, periods of mountain building, although they cannot reveal the amounts by which surfaces have been raised or lowered in an absolute frame of reference.
2. History and application of the basin-wide cosmogenic radionuclide analysis

2.1 Introduction:
Traditionally, denudation rates have been estimated from sediment yield studies operating over years or decades (e.g. Saunders and Young, 1983), from sediment volumes deposited in basins (e.g. Leeder, 1997; Mathews, 1975; Reneau et al., 1989), or by the depths to which dated surfaces have been incised (e.g. Ruxton and McDougall, 1967). The development of accelerator mass spectrometry produced instruments sensitive enough to measure trace concentrations of in-situ-produced cosmogenic radionuclides from mineral samples (Fifield, 1999; Tuniz et al., 1998). This has provided the geomorphologist with the ability to calculate denudation rates averaged over time scales which are more relevant to the operation of landforming processes (Cockburn and Summerfield, 2004). One variant of this technique involves sampling alluvial sediments to obtain the basin-averaged cosmogenic radionuclide concentration, allowing quantification of basin-wide denudation rates (herein referred to as the ‘basin-wide approach’).
In this chapter a history of the use of cosmogenic radionuclide analysis to derive basin-wide denudation rates will be presented. Firstly, the initial development and testing of the cosmogenic technique and basin-wide approach will be discussed. Secondly, studies which have utilised basin-wide cosmogenic radionuclide analysis will be discussed in relation to the results and methodologies used. These studies will be grouped according to their area of application.

2.2 Origins of the technique:

2.2.1 Denudation rates:

In the late 1970s accelerator mass spectrometry was developed in response to a need in the radiocarbon-dating community to be able to date samples smaller than the 1 g of carbon required by decay-counting techniques (Fifield, 1999). The possibility of deriving denudation rates from in-situ-produced cosmogenic radionuclide concentrations using accelerator mass spectrometry technology was first discussed by Lal and Arnold (1985). However, it was the seminal works of Nishiizumi et al. (1989) and Lal (1991) which provided the protocols for the measurement of $^{10}$Be and $^{26}$Al produced in-situ in quartz by cosmic radiation (or cosmic rays) and the theory required to convert these measurements to denudation rates and exposure ages. Although some of this initial work has been refined (Stone, 1999; 2000), these earlier publications provide the fundamental principles required for deriving denudation rates from cosmogenic radionuclide concentrations. Since these studies were published the growth of applications of cosmogenic radionuclide analysis is testament to its versatility and to the need which exists in the earth science
community to quantify denudation rates (Bierman et al., 2002; Cockburn and Summerfield, 2004).

2.2.2 Basin-wide denudation, initial work:

Three publications in the mid-1990s promoted the use of in-situ cosmogenic radionuclide concentrations in alluvial sediments to measure denudation rates, and provided the theoretical grounding and testing of the basin-wide approach. Brown et al. (1995) used soils, bedrock and alluvial sediment samples to investigate denudation in a small (3.26 km$^2$) Puerto Rico watershed. This work found an approximate agreement between the estimate of denudation over four years from constituent analysis of runoff solute concentrations (75±37 x 10$^{-3}$ mm/a), with that averaged over several thousand years from cosmogenic radionuclide concentrations in alluvial sediments (43±15 x 10$^{-3}$ mm/a). However, the different time scales over which rates were measured by this test precludes any confirmation that the technique is capable of measuring basin-wide denudation (Granger et al., 1996). The episodic mass movement processes dominant in the basin sampled were suggested to explain a bias of cosmogenic radionuclides found in differing sediment size fractions. The bias is the result of mass produced by shallow landslides mixing with the smaller size fractions, mobilised by sheetwash from more stable slopes. The landsliding material has a larger particle size but lower radionuclide concentration. In this way Brown et al. (1995) identified the operation of geomorphic processes and quantified their contribution to the mass flux from slopes. Granger et al. (1996) provided a more rigorous testing of the technique from catchments draining the Fort Sage Mountains onto the basin of the dried Lake Lahontan, California. They measured
the volumes of two alluvial fans deposited on the lake-bed and converted them into rates of denudation, using radiocarbon dates of the desiccation of the lake bed to estimate the period of time over which sediments had accumulated. The rates calculated for the two basins using this method were $58\pm 14 \times 10^{-3}$ and $30\pm 5 \times 10^{-3}$ mm/a. These rates were compared to those obtained using the cosmogenic radionuclide concentrations from the alluvial sediments leaving the basins which were $60\pm 14 \times 10^{-3}$ and $36\pm 9 \times 10^{-3}$ mm/a, respectively. The close agreement between the rates from the different techniques, measured over comparable time scales, suggests the cosmogenic radionuclide method is a viable one for the calculation of basin-wide denudation rates. Granger et al. (1996) use the denudation rates derived in a comparison with basin slope and proposed an exponential increase in denudation with increases in hillslope gradient. Although Brown et al. (1995) and Granger et al. (1996) considered the theoretical assumptions behind the basin-wide approach, they were most comprehensively addressed by Bierman and Steig (1996) who modelled the cosmogenic radionuclide inventory of a catchment. Their treatment showed that, providing sediments leaving a basin are thoroughly mixed, sampling channel alluvium gives a spatially representative, averaged denudation rate for the catchment area upstream of the sample site. These concepts will be considered more fully in Chapters 3 and 4. Bierman and Steig (1996) also investigated the affect that episodic denudational events of varying magnitudes might have on the cosmogenic radionuclide concentration of alluvial sediments. By modelling denudation rates which fluctuate over time according to a step function, they were able to show that changes in denudation rates in rapidly denuding basins can produce the greatest error when sampled using the basin-wide approach.
However, after a perturbation, rapidly denuding basins are quicker to attain a new secular equilibrium with respect to cosmogenic radionuclide concentration. Hence, the potential error will be minimised most quickly after any change in denudation rate if the rates are rapid.

2.3 Basin-wide denudation applications:

After this initial testing came the application of the basin-wide approach and three main areas of research can be identified: 1. investigation of basin-scale sedimentary processes; 2. investigation of the controls of denudation rates and 3. landscape evolution studies. These areas are discussed individually below, although several studies present results that are relevant to more than one of these groupings.

In sediment process studies utilising cosmogenic radionuclide concentrations several sampling techniques are often applied simultaneously, with many studies using cosmogenic radionuclide concentrations of bedrock and detrital samples to quantify the processes of soil production, mineral enrichment and sediment transport (e.g. Heimsath et al., 1997; 2000; Kyle et al., 2002; Phillips et al., 1998; Riebe et al., 2001a; 2001b; Small and Anderson, 1999). However, the focus of subsequent chapters is on utilising the basin-wide approach and so only studies which have used cosmogenic radionuclide concentrations from alluvium specifically to derive basin-averaged denudation rates have been considered in detail.
2.3.1 Sedimentary processes:

Clapp et al. (2000) published one of the first studies to employ the basin-wide approach in order to quantify sediment transport. In a small hyper-arid drainage basin in southern Israel they utilised modern sediment yield data and compared it with the intermediate-term cosmogenic rates from a mixture of 33 paired $^{10}$Be and $^{26}$Al cosmogenic alluvial, colluvial and bedrock samples. The suggestion from this data is that the increase in denudation rate of between $42 \times 10^{-3}$ and $51 \times 10^{-3}$ mm/a, derived over 33 years by sediment yields, in comparison to the cosmogenic rates of $\sim 27 \times 10^{-3}$ mm/a, derived over several tens of thousands of years, indicates present day excavation of stored sediment. Clapp et al. (2000) found that cosmogenic radionuclide concentrations can be employed as source tracers allowing them to suggest the majority of sediment leaving the basin was hillslope colluvium. The mixed lithology of the basin sampled allowed the influence of rock-type on radionuclide concentration to be investigated and yielded no significant difference in denudation rates between the gneissic granite, schist and amphibolites underlying the basin. Clapp et al. (2001; 2002) extended the work in arid basin settings using cosmogenic radionuclide concentrations in Arizona and New Mexico basins in the south-western United States to investigate issues of sediment production, storage, mixing and transport. The work in Arizona (Clapp et al., 2002) focused on a mesoscale ($187 \text{ km}^2$) basin which contained a broad, braided ephemeral channel. Fourteen alluvial sample sites were identified in this channel while a further 13 sample sites consisting of a mixture of bedrock, colluvium and alluvium, were allocated to a small sub-basin off the main channel. Several aliquots of different grain size were collected from each sample site. They recorded a lack of
cosmogenic radionuclide concentration bias in different grain sizes which was attributed to the thorough mixing of sediment during transport. This work further showed that the denudation rates of the small, more mountainous, sub-basin ($\sim 26 \times 10^{-3}$ mm/a) were less than the denudation rates from the larger trunk channel ($\sim 38 \times 10^{-3}$ mm/a) suggesting that material is presently leaving the basin more rapidly than it is being produced. Hillslope colluvial sources were found to provide the majority of mass to the main channel while transport of sediments downstream was surmised to occur in pulses on the basis of radionuclide concentrations found in colluvial, alluvial and exposed bedrock. A sediment mixing-model was constructed using the cosmogenic radionuclide concentrations from different parts of the basin as ‘signatures’ to infer the ratio of sediment mixing which was occurring at tributary junctions. Clapp et al. (2001) sampled a $\sim 17$ km$^2$ low relief basin in a semi-arid setting in New Mexico. They found no grain size dependence of radionuclide concentrations from the seven alluvial samples they collected. Denudation rates derived from three bedrock samples indicates that bedrock weathers more rapidly beneath a cover of colluvium. Two depth profiles were also sampled and, along with the radionuclide concentrations from the channels, showed that the basin is in a steady-state with respect to production and removal of sediment. The cosmogenically derived denudation rate of $10^{-2}$ mm/a over the last several thousand years is similar to the modern rate of $15^{-2}$ mm/a from a study over two years using sediment traps.

Kirchner et al. (2001) considered the transport of sediment from forested catchments in Idaho ranging in size from $10^1$ to $10^5$ km$^2$. They compared denudation rates over
10^1 year time scales from sediment fluxes with rates derived over 10^4 years and 10^7 years from 32 cosmogenic radionuclide alluvium samples and fission-track data, respectively. The average cosmogenic and fission-track derived denudation rates were similar, between ~50x10^{-3} to 100x10^{-3} mm/a. The sediment yield results, however, were on average ~17-fold less than the longer-term estimates. These results were interpreted as indicating high magnitude, low frequency events giving rise to an episodic regime of sediment transport from mountainous catchments. The episodic nature of significant sediment transport events was not recorded by the short term monitoring of sediment loads but is evident in the longer-term, cosmogenic and fission-track, measurements.

Perg et al. (2003) utilised the cosmogenic radionuclide concentrations of beach deposits in order to consider the source areas of the littoral cell. Denudation rates derived from basins draining into the littoral cell of a stretch of Santa Cruz coastline, northern California, ranged between 80±10x10^{-3} and 28±20x10^{-3} mm/a. The cliffs along this stretch of coast are topped by palaeobeach deposits and are back-wearing at ~100 mm/a, depositing the high concentration palaeobeach sediments into the littoral cell. Sampling channel alluvium from five basins, 60 to 319 km^2 in size, before they reach the coast, and also the palaeobeach sediments on the cliffs, provided end member values from which the degree of mixing in the littoral cell could be derived. The radionuclide concentrations of 19 beach sand samples were used to estimate source areas and the amount of drift which occurs along this stretch of coast. Along with the studies of Clapp et al. (2000; 2001; 2002) and Kirchner et al. (2001) this work highlights one of the fundamentals of the basin-wide approach,
advocated by Bierman and Steig (1996), in that the sediments sampled must be thoroughly mixed in order to obtain a representative radionuclide concentration from the upstream area (see Chapter 4).

2.3.2 Controls of denudation:

Granger et al. (2001) and Riebe et al. (2000; 2001c) built upon the work by Granger et al. (1996) and used cosmogenic radionuclide concentrations in alluvial sediments from a variety of basin settings in the Sierra Nevada to investigate denudation and weathering in relation to factors thought to control rates. Granger et al. (2001) used 25 drainage basin samples, measuring both $^{10}$Be and $^{26}$Al, to derive denudation rates of between $15 \times 10^{-3}$ and $60 \times 10^{-3}$ mm/a. They also collected several rock samples. Interpretation of these results indicated that boulder armouring of slopes modulates denudation and leads to lower denudation rates. The similarity of denudation rates from both steep and gentle topography is considered to represent uniform lowering of the landscape and is facilitated by boulder armouring. In this study Granger et al. (2001) discussed the importance of including muogenic production in denudation rate studies and provided the formula and constants required to do so, a topic discussed further in Chapter 3. Riebe et al. (2000) used the denudation rates from 56 samples across seven different sites in the Sierra Nevada, California, incorporating the samples of Granger et al. (1996; 2001), to investigate the controls of denudation in non-glacial basins. Results showed that hillslope gradient does not appear to control denudation rates as they are uniform across catchments with different slopes. However, a correlation was found between slope and denudation rate in those basins close to an incising trunk stream or fault scarp, suggesting that
the presence of local base-level drop is driving slope and denudation rate increases. Riebe et al. (2001c) used the same samples measured by their earlier study (2000) and compared the results with climatic data. The nine-fold and four-fold variations in precipitation and temperature were found not to be related to the denudation rates, which ranged between $24\pm5\times10^{-3}$ and $61\pm4\times10^{-3}$ mm/a. However, it should be noted that these denudation rates are averages of several samples taken from each site which, individually, show up to nine-fold variation. The implications of this work are that over the climatic range of the environments sampled denudation rates appear to be insensitive to climate, although only the amount not the seasonality of precipitation was considered. The absence of a climatic influence on denudation rates in the Sierra Nevada recorded by Riebe et al. (2001c) compliments the work of Bierman and Caffee (2001), who sampled cosmogenic radionuclides across the Namib Desert from the coast, over the escarpment to the highland region to highlight a lack of correlation between denudation rates and mean annual precipitation. Sixty-six paired $^{10}$Be and $^{26}$Al samples consisting of seven alluvial sediment samples from basins ranging in size from $10^0$ to $10^5$ km$^2$ were collected. Denudation rates from the alluvial sediment samples range from $3\times10^{-3}$ to $16\times10^{-3}$ mm/a, showing on average slightly higher denudation rates than the 47 bedrock outcrop samples measured. Across the sampling transect there is an increase in mean annual precipitation from $<50$ mm/a to $>400$ mm/a which is not reflected in the pattern of denudation rates, although a variation would probably not have been expected over such a range of low precipitation values.
The cosmogenic radionuclide technique averages denudation over sufficient periods of time that results are typically insensitive to anthropogenic influence and can provide a benchmark rate against which any human impact can be observed. Several studies have taken advantage of the basin-wide approach to obtain both spatially and temporally averaged base-rates of denudation for comparison with modern sediment yields. Brown et al. (1998) measured the $^{10}$Be concentration from a total of 14 alluvial samples and five samples of landslide material from two basins in Puerto Rico, incorporating a further nine samples from an earlier study (Brown et al., 1995). Of the two basins, one had been impacted by humans and the other not; denudation rates of $\sim 85 \times 10^{-3}$ mm/a and $\sim 43 \times 10^{-3}$ mm/a, respectively, were recorded. A bias of radionuclide concentration with grain size was observed and considered to result from the mobilisation of coarse, low concentration material from depth by shallow landsliding (section 2.3.1). Comparing the cosmogenic derived denudation rates with modern sediment yield studies shows that the cosmogenic rate of $85 \times 10^{-3}$ mm/a is much less than the short-term rate of $75 \times 10^{-2}$ mm/a, recorded over four years. Brown et al. (1998) argued that anthropogenic influence has increased denudation rates in the former which was once similar in rate to the undisturbed catchment. Hewawasam et al. (2003) also concluded there had been an increase in modern denudation rates in comparison to the ‘natural’, or background, rates deduced over thousand year timescales from cosmogenic radionuclides in a Sri Lankan drainage basin. In this study eight small basins ($< 1 \text{ km}^2$), and nineteen larger basins (9-844 km$^2$) were sampled by collecting alluvial sediments at the mouths of the catchments. Cosmogenic derived denudation rates range from $5 \times 10^{-3}$ to $45 \times 10^{-3}$ mm/a, averaged over the past 19 to 147 ka. Short-term sediment loads
measured over the last three to thirty years are 10 to 100-fold higher than the cosmogenic rates. This increase is attributed to recent land use and deforestation which is destabilizing slopes.

Schaller *et al.* (2001) reported a decrease in modern sediment yields in comparison to the longer cosmogenic time scales. Schaller *et al.* (2001) presented a denudation rate decrease since between 10 to 40 ka ago based on the alluvial sampling of four large middle-European drainage basins and their sub-basins (10⁵-10⁹ km²) with the modern bedload measurements. Seventy-two ¹⁰Be samples were processed and produced denudation rates ranging from 112±14x10⁻³ to 5.0±0.3x10⁻³ mm/a. The decrease in modern rates is assigned to either underestimation of modern sediment loads because the period of measurement fails to reflect low-frequency high-magnitude events, or due to release of stored sediments. Alternatively, higher Pleistocene denudation rates cannot be discounted. Furthermore, Schaller *et al.* (2001) provided a methodology for dealing with muogenic production of cosmogenic radionuclides in denudation rate calculations. Schaller *et al.* (2002) used the same terrace sediment samples measured by Schaller *et al.* (2001) to investigate rates of basin-wide denudation at the macroscale. By sampling independently dated alluvial terraces they were able to constrain rates of denudation over different periods of time from the last 30 ka. A decrease of between 4 to 1.5-fold in denudation rates since the early Holocene was recorded and proposed to be the result of a change from glacial to post-glacial conditions. During the Last Glacial Maximum, periglacial conditions would have existed over large areas of the catchments sampled and this is proposed as the reason for the higher denudation
rates. Integration of the higher Pleistocene denudation rates over the Holocene is suggested as the reason why Schaller et al. (2001) recorded high background denudation rates using the cosmogenic technique.

These studies, conducted over different scales, raise an interesting series of questions. What timespan of measurement is appropriate to capture the human impact on denudation rates? At what spatial and temporal scales are other controls of denudation such as climate, relief and lithology effective? Are measurements of denudation rates averaged over different temporal and spatial scales recording the same phenomena and, if not, is it appropriate to compare them? These questions will be reconsidered in Chapter 8 in light of the work presented here.

2.3.3 Landscape evolution:
Landscape evolution studies have benefited from a range of cosmogenic radionuclide applications allowing the rates of denudation, critical for describing the pace and processes of topographic development, to be quantified. Such applications have included dating of strath terraces, or alluvial sediment deposited in caves, above modern channels to infer the tempo of orogenic development in tectonically active regions (Burbank et al., 1996; Pratt et al., 2002; Stock et al., 2004). Sampling well mixed alluvial sediments provides a spatially integrated measure of cosmogenic radionuclide concentrations across a basin, making it particularly relevant to mesoscale and macroscale geomorphological study. However, there are few studies employing the basin-wide approach in order to investigate the landform evolution of tectonically active mountainous regions. This may be partly due to the potential of
invalidating assumptions of mixing required by the basin-wide cosmogenic approach (section 4.5.3). Alternatively it may be because rapidly denuding regions are subject to larger errors, making results less conclusive (section 8.3.2.1). However, if such problems can be resolved, cosmogenic radionuclide concentrations from mountainous catchments present a method by which to constrain and test long-standing models of orogenesis.

One study has provided rates from a rapidly denuding part of the Himalaya (Vance et al., 2003). Fifteen samples were collected from $10^3$ km$^2$ basins, recording denudation rates of between 2.7 and 1.3 mm/a. Vance et al. (2003) found mass balance studies and fission-track data provide similar results but that high temperature $^{40}$Ar/$^{39}$Ar thermochronometry in muscovite suggests lower denudation rates. Combined, the data suggested an increase in denudation occurred several million years ago and that since then this high rate has been uniform and is observed over a range of temporal scales. Furthermore, a log-linear relationship between denudation and relief was found to be present even when the Himalayan data is combined with the denudation rate data of Schaller et al. (2001) from lowland middle-Europe. Matmon et al. (2003) provided a study of the cosmogenic radionuclide inventories of mountainous catchments in the more slowly denuding Great Smoky Mountains at the southern end of the Appalachian Mountains. Twenty-six alluvial samples from a 340 km$^2$ basin, divided into progressively smaller sub-basins, underlain by a mixture of metamorphic sedimentary and gneiss lithologies, gave denudation rates ranging from $37 \times 10^{-3}$ to $17 \times 10^{-3}$ mm/a. The greatest variance between denudation rates is found in the small scale ($10^0$ km$^2$)
sub-basins. When averaged, the denudation rates from the sub-basins approximated the rates from the larger basins. The suggestion from this is that downstream alluvial sediments are homogenised sufficiently for denudation rate differences between the small scale basins to be averaged when sampling larger basins. Utilising published fission-track and sediment yield data showed denudation rates have been similar over $10^0$ to $10^6$ a time scales and from these results Matmon et al. (2003) proposed a temporal and spatial homogeneity of denudation rates in these mountains. Furthermore, Matmon et al. (2003) discussed the mixing of sediments in the basins sampled by employing a model similar to the one used by Clapp et al. (2002).

A question raised by all the basin-wide approach studies presented in this chapter is how much variation in rates, whether temporally or spatially, should be considered as significant? Vance et al. (2003) discussed their evidence for an increased rate on the basis of a three to six-fold increase between rates measured by different techniques. While the average ~17-fold difference in denudation rates measured over different time scales by Kirchner et al. (2001) presented a convincing argument of change, the 4 to 1.5-fold differences presented by Clapp et al. (2000), Bierman et al. (2001) and Schaller et al. (2001) were less so, whilst Matmon et al. (2003) interpreted differences of this magnitude as being indicative of a temporal homogeneity. Clearly what constitutes a significant change in rates through time is dependent on the type of environment sampled but has yet to be universally defined and leaves results open to ambiguous interpretation. Accordingly this topic will be considered further in Chapter 8, in light of the denudation rate results obtained by this study.
2.4 Summary:

There has been growth in both the use and applications of cosmogenic radionuclides in alluvial sediments since the method was first suggested by Lal and Arnold (1985). It is a technique which allows, for the first time, a readily applicable method for deriving spatially integrated denudation rates over $10^2$-$10^6$ year time scales. Many of the problematic issues relating to the application of technique and the assumptions it requires have been dealt with by studies using the basin-wide approach in new environments. However, little application has been undertaken in rapidly denuding, tectonically active, mountains. As with any technique, it entails a particular set of assumptions which require clarification and these will be addressed next in Chapter 3 as the calculations, chemistry and sampling protocols required to derive denudation rates from alluvial sediments in rapidly denuding environments are discussed.
3. Calculation of basin-wide denudation from cosmogenic radionuclides in alluvial sediment

3.1 Introduction:

Cosmic rays, composed of lower energy particles from the sun and higher energy particles from galactic sources, continually bombard the earth from space. Upon penetrating the earth’s surface the cosmic rays are attenuated by collisions with other nuclei, resulting in the production of \textit{in-situ} cosmogenic radionuclides. Concentrations of cosmogenic radionuclides at the earth’s surface will, therefore, be a function of the rate of production and the residence time of bedrock at the surface prior to removal by denudation (Lal 1991). By calculating a rate of cosmogenic radionuclide production and being able to measure, by accelerator mass spectrometry, the \textit{in-situ}-produced concentration in mineral samples, the rate of denudation can be derived. Fundamental to an accurate assessment of the rate of denudation is an understanding of the mechanisms and rates at which cosmogenic radionuclides are produced. It is appropriate, therefore, to introduce cosmogenic radionuclide production systematics prior to discussing how they are used to model
denudation. As such, a discussion of the principles of cosmic radiation and the
different mechanisms by which cosmogenic radionuclides are produced will be
given first. Secondly, the assumptions and formulations employed to derive
denudation rates using cosmogenic radionuclides will be considered, followed by the
various corrections which have to be made for geomagnetic field, latitude,
averaging time of the technique before the last section which will discuss the
atmospheric depth and shielding. The utility of digital elevation models to meet
these required corrections will be shown. There will be a discussion concerning the
protocols and chemistry behind the sample collection, processing and measurement
adopted in this study. Only cosmogenic $^{10}\text{Be}$ is used to determine denudation rates
in this study and so the discussion will primarily reflect its application. However,
because they are often considered together, the production corrections for both
cosmogenic $^{10}\text{Be}$ and $^{26}\text{Al}$ will be presented. Moreover, many of the corrections
applied in cosmogenic radionuclide analysis are specific to location and so where
appropriate the coordinates of the San Bernardino Mountains, California, have been
used.

3.2 Cosmogenic radionuclide production systematics:

3.2.1 Cosmic radiation:
The majority of cosmic radiation pertinent to the production of cosmogenic
radionuclides originates from within the Milky Way galaxy (Gosse and Phillips,
2001). These galactic-cosmic-rays are composed of energetic protons (~87%), alpha
particles (~12%) and a few heavier nuclei (~1%) (Masarik and Reedy, 1995). Other
ccontributions from the sun, termed solar-cosmic-rays, are confined to the lower end
of the energy spectra ($<10^9$ eV). These usually only penetrate the earth’s geomagnetic field at latitudes above 60° producing cosmogenic radionuclides just in the very top of the atmosphere (Lal, 2000; Masarik and Reedy, 1995). However, higher than average proton fluxes resulting from solar-particle events may make a more significant contribution (Masarik and Reedy, 1995). Some high-energy cosmic radiation has been suggested to originate from extra-galactic sources (Friedlander, 1989).

Cosmic radiation passes through several ‘filters’ on its passage to the earth’s surface, namely the heliosphere, the earth’s geomagnetic field, and the atmosphere (Lal, 2000). The heliosphere extends beyond the solar system and relates to the outward diffusion of plasma from the sun, hence it has the ability to modulate the galactic-cosmic-ray flux depending on temporal variations in solar activity (Lal, 2000; Masarik and Reedy, 1995). However, for cosmogenic radionuclide studies at the earth’s surface, more attention has been paid to geomagnetic field variations and atmospheric effects.

Cosmic radiation is comprised of mostly charged protons and so is deflected as it passes through the earth’s geomagnetic field. Whether or not a particular particle will pass through is dependent on its momentum to charge ratio and the inclination of the field (Gosse and Phillips, 2001). The momentum to charge ratio of particles is termed the cutoff-rigidity and as such it defines the minimum energy incoming particles must have to avoid deflection by the earth’s geomagnetic field. The higher the angle of incidence between the earth’s geomagnetic field and incoming
radiation, the higher the cutoff-rigidity required to avoid deflection. Hence, at
equatorial latitudes, where the field is perpendicular to the average incidence angle
of cosmic radiation, the radiation which reaches the atmosphere will tend to be
composed of higher energy particles than at the poles where the magnetic field will
permit radiation of lower energies (Dunai, 2000; Gosse and Phillips, 2001). Above
the 60° latitude ‘knee’, the cutoff-rigidity of the field has dropped to 0.38 GeV (Lal,
2000). This value is below the minimum energy of protons in cosmic radiation
which results in a uniform cosmic ray flux reaching the atmosphere between 60° and
the poles (Dunai, 2000). Geomagnetic variations, such as changes in
palaeointensity and dipole movement, can influence cosmic radiation and must be
considered when deriving production rates for cosmogenic radionuclides. This will
be discussed further in section 3.4.2.1.

The final ‘filter’ cosmic radiation will experience before it reaches the earth’s
surface is the atmosphere. Prior to entering the earth’s atmosphere, cosmic rays are
termed primary. As primary cosmic radiation collides with gas particles in the
atmosphere a cascade of secondary protons, neutrons, pions and other strongly
interacting particles, the so-called hadronic cascade, is produced (Masarik and
Reedy, 1995). This is often termed secondary cosmic radiation. Neutrons comprise
the greatest component of the secondary cosmic radiation with protons making up
<10% (Nishiizumi et al., 1989), while muons are produced by the decay of pions
near the top of the atmosphere (Stone et al., 1998a). The transition from mostly
protons at the top of the atmosphere to mostly neutrons at the earth’s surface results
from the higher chance of neutrons being emitted during nuclear reactions in the
hadronic cascade and because they have no charge, neutrons have less chance of further particle interaction (Gosse and Phillips, 2001). It is the neutrons and muons in secondary cosmic radiation which are responsible for in-situ-production of $^{10}$Be and $^{26}$Al in bedrock at the earth’s surface. Cosmogenic radionuclides produced in the atmosphere during the cascade include $^{10}$Be (often termed ‘meteoric’ or ‘garden-variety’) which has been used in several studies of soil processes (e.g. Pavich et al., 1984; Monaghan et al., 1992) but should not be confused with $^{10}$Be produced in-situ within a mineral lattice. $^{26}$Al is also produced in the atmosphere but not to the same degree as $^{10}$Be (Gosse and Phillips, 2001).

Attenuation of this cascade of secondary particles is dependant on atmospheric pressure, hence the production of in-situ-produced cosmogenic radionuclides must be scaled for altitude (Lal, 1991; Stone, 2000). Furthermore, in comparison to neutrons, muons are attenuated less steeply in the atmosphere and so should be scaled independently (Gosse and Phillips, 2001; Heisinger and Nolte, 2000; Stone, 2000). After passing through the atmosphere, the depth to which secondary cosmic radiation is able to penetrate into rock at the earth’s surface is quantified by the absorption coefficient (Lal, 1991). This variable is inversely proportional to the attenuation length of the cosmic rays and proportional to the rock density. The concept of the attenuation length is not intuitive but can be described as the weight of atoms, in grams, which would be measured along a 1 cm$^2$ cross sectional trajectory through a medium, in this case rock, until the cosmic-ray-flux intensity is reduced to a factor of 1/e (i.e. to ~37%) (Friedlander, 1989). As such, radiation with
a higher energy, or which interacts less with other particles, will have a higher attenuation length and penetrate more deeply.

3.2.1.1 Spallation:

The fast, or energetic, neutron component of secondary cosmic radiation is able to break apart target atoms in the lithosphere to create cosmogenic $^{10}$Be and $^{26}$Al in a spallation reaction. Energy will be lost from these neutrons by successive reactions till they reach $\sim$1-5 MeV whereupon they are no longer able to cause spallation and their energy will dissipate till they reach thermal neutron status, $\sim$0.1 MeV-0.5 eV, which is important for the production of cosmogenic $^{36}$Cl and $^{41}$Ca but not $^{10}$Be or $^{26}$Al (Gosse and Phillips, 2001). Quartz is the typical target mineral for in-situ-produced $^{10}$Be and $^{26}$Al in cosmogenic radionuclide studies due to its abundance and favourable chemical properties (Bierman et al., 2002). Typically, although not exclusively, $^{10}$Be will result from spallation of $^{16}$O (n,4p3n)$^1$ and $^{26}$Al from $^{28}$Si (n,2pn) (Nishiizumi et al., 1989). Spallation reactions dominate at the earth’s surface contributing $\sim$97% of the total $^{10}$Be and $^{26}$Al concentrations at sea level (Stone et al., 1998b; Stone, 2000). However, with an attenuation length of $\sim$160 g/cm$^2$ spallation reactions will decrease exponentially in rock such that spallation becomes relatively unimportant a few metres below the surface in comparison to muogenic production (Heisinger and Nolte, 2000).

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1The values in brackets denote nuclei involved in reactions where n and p are neutrons and protons.
3.2.1.2 Muogenic production:

Muogenic production, whilst insignificant in comparison to spallation for many cosmogenic radionuclide applications, must be considered when deriving denudation rates (Granger et al., 2001; Heisinger and Nolte, 2000; Heisinger et al., 2002a; section 3.3.3). Slow negative muons ($\mu^-$) result from the decay of negative pions and decay rapidly themselves with a half-life of $\sim 2 \times 10^{-6}$ seconds but may be captured by a charged nucleus which becomes excited and emits neutrinos and nucleons (Gosse and Phillips, 2001; Heisinger et al., 2002a). In the case of production of $^{10}$Be in quartz the nuclear reaction is typically $^{16}$O ($\mu^-,3p3n$) and for $^{26}$Al in quartz $^{28}$Si ($\mu^-,2n$) (Nishiizumi et al., 1989). Deceleration of fast muons ($\mu$) produces a shower of various types of secondary particle which can produce cosmogenic radionuclides (Heisinger et al., 2002b; Stone et al., 1998a). Muons occupy an intermediate niche between highly reactive (e.g. neutrons and protons) and non-reacting (e.g. neutrino) particles and consequently are attenuated less than the neutron component of the secondary cosmic flux (Gosse and Phillips, 2001). The depth to which they can penetrate into bedrock is also a function of their energy and does not approximate as well to an exponential as production by spallation; furthermore, fast muons will attenuate more gradually than slow muons requiring separate scaling for each (Granger et al., 2001; Stone et al., 1998a). An average attenuation length for muons is $\sim 1500$ g/cm$^2$, about ten-fold greater than for spallation (Gosse and Phillips, 2001).
3.2.1.3 Non-cosmogenic production of $^{10}$Be and $^{26}$Al:

$^{10}$Be can also be produced in minerals by interaction with the decay products of Li, U and Th but except in lithologies containing unusually high amounts of these radiogenic isotopes, the amount will be negligible (Brown et al., 1991). Radiogenic decay of U and Th to Na can produce appreciable background concentrations of $^{26}$Al (Gosse and Phillips, 2001; Sharma and Middleton, 1989). Because U is concentrated in non-quartz minerals, however, this effect may be reduced. The non-cosmogenic production of $^{10}$Be and $^{26}$Al can probably, therefore, be considered negligible when using quartz samples in most lithologies (Bierman, 1994).

3.3 Model Theory:

3.3.1 Denudation rate model:

Knowledge of cosmogenic radionuclide production mechanisms now allows the discussion to turn to the theoretical underpinnings behind the models of denudation. Lal and Arnold (1985) first discussed the use of $^{10}$Be and $^{26}$Al produced in-situ in quartz for the purpose of denudation rate estimates. However, it was Lal’s seminal 1991 paper which developed the production rate scaling required for the calculation of mass loss using cosmogenic radionuclide concentrations based on equation [3.1].

$$N(x,t) = N(x,0)e^{-\lambda t} + \frac{P(x)}{\lambda + \mu \varepsilon} e^{-\mu \varepsilon} (1 - e^{-(\lambda + \mu \varepsilon) t})$$

[3.1]
In this formulation the cosmogenic radionuclide concentration $N$ (atoms/g) in a bedrock surface at depth $x$ (cm) and time $t$ (years) is a function of the initial cosmogenic radionuclide concentration ($t=0$) and the production rate $P$ (atoms/g/a), minus the loss by radiogenic decay $\lambda$ (a$^{-1}$) and denudation rate $\varepsilon$ (cm/a). The absorption coefficient $\mu$ (cm$^{-1}$) is a measure of the cosmogenic radionuclide attenuation which occurs in matter and is a function of the density $\rho$ (g/cm$^3$) over the attenuation length $\Lambda$ (g/cm$^2$); $\mu=\rho/\Lambda$ (section 3.2.1). An alternative expression for this is as the coefficient of penetration, $z^*$, which is the depth in centimetres that cosmic radiation will reach before the intensity drops to a factor of 1/e. This can be expressed as $z^* = \Lambda/\rho$ (i.e. the inverse of the absorption coefficient).

It is assumed in formula [3.1] that the production rate has not varied significantly through time (section 3.4.2.1). The equation can be simplified by making further assumptions. Firstly, that the rate of denudation has been consistent enough for a sufficient period of time to allow the cosmogenic radionuclide concentration of a surface to have achieved secular equilibrium with respect to production by incoming cosmic radiation and loss through decay and denudation (Fig. 3.1, section 3.5). This essentially makes time redundant in equation [3.1] and it can be considered to tend towards infinity. Secondly, the assumptions are made that the minerals at the surface have had no prior exposure to cosmic radiation before denudation of the overlying mass brought them into the penetrative range of cosmic radiation and that denudation is occurring at the surface (i.e. $x=0$). Employing these assumptions gives equation [3.2] where $P(0)$ and $N(0)$ are the production and cosmogenic radionuclide concentration at the surface.
\[ N(o) = \frac{P(o)}{\lambda + \mu \varepsilon} \]

[3.2]

Rearranging this to solve for denudation and employing the relationship that \(1/\mu = z^*\) (see above) gives equation [3.3].

\[ \varepsilon = \left[ \frac{P(o) - \lambda}{N(o)} \right] \cdot z^* \]

[3.3]

3.3.2 Basin-wide denudation rate model:

Bierman and Steig (1996) presented a reservoir model in which cosmogenic radionuclide concentrations are measured in alluvial sediment as opposed to an intact bedrock surface giving spatially averaged, or basin-wide, rates of denudation (Fig. 3.2). They note that by integrating equation [3.2] the isotope inventory of a drainage basin can be calculated and from this the basin-wide denudation. However, several caveats apply. Firstly, the assumptions introduced by Lal (1991) for the use of equation [3.2] must also be met when using the basin-wide approach. Therefore, the basin averaged production rate is assumed to be known and to have been constant and that the integrated denudation of the basin has been temporally consistent, though not necessarily spatially homogeneous, for long enough such that the basin has achieved a steady-state with respect to cosmogenic radionuclide
Figure 3.1 From an initial cosmogenic radionuclide (CRN) concentration of zero, the build up of nuclides over time will reach a secular equilibrium between radionuclides produced and those lost to decay and denudation. For more rapid denudation, the time to achieve equilibrium is shorter.

Figure 3.2 The reservoir model of cosmogenic radionuclide concentration (adapted from Bierman and Steig 1996). The concentration in the basin ‘reservoir’ is a function of the rate of input by cosmogenic radiation and the rate of output by denudation. Radiogenic input by U and Th decay is assumed negligible, as may be the output by radioactive decay (see text for discussion).
concentration (Fig. 3.1). Secondly, further assumptions must be made which are: 1. sediment sampled is thoroughly mixed and hence representative of the entire area upstream of the sampled point; 2. denudation is occurring primarily by incremental mass removal and not deep-seated landsliding and that; 3. the mineral from which the cosmogenic radionuclides will be extracted is distributed uniformly throughout the basin. Assumptions 2 and 3 can be assessed by field observation, whereas assumption 1 is more problematic and will be considered further in Chapter 4.

Bierman and Steig (1996) modelled the issue of spatially heterogeneous denudation rates in a basin. They note that for an idealised basin eroding at an average rate $\bar{\varepsilon}$, which is composed of multiple sub-basins eroding at different rates, $\varepsilon_1$ and $\varepsilon_2$, the average cosmogenic radionuclide concentration resulting from $\bar{\varepsilon}$ will not equal the integrated sub-basin concentrations. This is shown by equation [3.4].

$$\frac{P_1(o)}{\lambda + \mu \varepsilon_1} + \frac{P_2(o)}{\lambda + \mu \varepsilon_2} \neq \frac{\bar{P}(o)}{\lambda + \mu \bar{\varepsilon}}$$

[3.4]

The reason for this discrepancy is the variable influence of the decay constant, $\lambda$, with denudation. Cosmogenic $^{10}$Be and $^{26}$Al undergo radioactive decay with respective half lives of 1.52 Ma and 0.71 Ma (Gosse and Phillips, 2001). The influence the decay constant has on the cosmogenic radionuclide concentrations of minerals at the earth’s surface is a function of how long that mineral has been subject to cosmic radiation, which depends on the rate at which it is approaching the
surface (i.e. the rate of denudation) and the penetration depth (or attenuation length) of the radiation. If denudation is low, minerals will take longer to reach the surface and more cosmogenic radionuclides will decay than if denudation was rapid. As decay is a non-linear process it has variable influence in each sub-basin and so cannot be represented by simple mixing, as shown in equation [3.4]. However, the relatively long half lives of $^{10}\text{Be}$ and $^{26}\text{Al}$ mean that for denudation rates greater than several metres per million years and sediment storage times less than several thousand years the loss by decay is negligible and the decay term can be ignored (Bierman and Steig, 1996). This resolves the discrepancy illustrated in [3.4] and results in equation [3.5].

$$\frac{P_1(o)}{\mu \varepsilon_1} + \frac{P_2(o)}{\mu \varepsilon_2} = \frac{\bar{P}(o)}{\mu \bar{\varepsilon}}$$

[3.5]

Note this also illustrates that results are not influenced by the spatial variability in denudation rates in relation to variability in production rates. That is to say, there will be no error introduced if the majority of sediment comes from an area of relatively high or low production within the basin because the nuclide concentration in sediment leaving the basin is a function of the average production over the average denudation (Bierman and Steig, 1996; Vance et al., 2003).
3.3.3 Inclusion of muogenic production:

In environments where denudation rates are low, production of $^{10}$Be and $^{26}$Al at the surface is dominated by spallation resulting from interactions with the hadronic component of cosmic radiation (Masarik and Reedy, 1995; Stone, 2000). As denudation rates increase, the proportion of cosmogenic $^{10}$Be and $^{26}$Al produced through muogenic interactions will also increase (Granger et al., 2001; Heisinger and Nolte, 2000; Heisinger et al., 2002a). This is because the muogenic component is attenuated less and so penetrates deeper than the nucleon component, hence, when denudation is low much of the muogenic produced $^{10}$Be and $^{26}$Al will have decayed before reaching the earth’s surface. However, as denudation rates increase, the time available for cosmogenic radionuclides to decay before reaching the surface decreases and so the relative contribution of cosmogenic radionuclides produced by fast and slow muogenic interactions rises. Granger et al. (2001) and Schaller et al. (2001) have described models to derive denudation rates from cosmogenic $^{10}$Be and $^{26}$Al incorporating muogenic production. Here, the Granger et al. (2001) model is employed due to its relative simplicity, as shown below in [3.6].

$$
N(\phi) = \left[ \frac{P(\phi)}{\lambda + \rho e / \Lambda} \right] + \left[ \frac{Y A_1}{\lambda + \rho e / L_1} \right] + \left[ \frac{Y A_2}{\lambda + \rho e / L_2} \right] + \left[ \frac{B}{\lambda + \rho e / L_3} \right]
$$

[3.6]

This formulation now incorporates a range of production rate values for cosmogenic $^{10}$Be and $^{26}$Al by slow muogenic capture ($Y A_1$ and $Y A_2$), where $Y$ is the yield per
stopped negative muon and $A_1$ and $A_2$ are production constants (muons/g/a), and by fast muon production $B$ (atoms/g/a). Respective attenuation lengths are included for slow muon capture, $L_1$ and $L_2$, and fast muon production, $L_3$. In Figure 3.3, equation [3.6] has been used to plot the relative contribution of spallation and fast and slow muons to the $^{10}$Be concentration at the earth’s surface across a range of denudation rates. This shows that at rates of denudation on the order of, or greater than, $10^{-3}$ mm/a the muogenic contribution becomes significant (~20% of the total), justifying the use of equation [3.6] in areas likely to have high denudation rates. When denudation rates are on the order of, or less than, $10^{-4}$ mm/a, muogenic production itself becomes such a small fraction of the total production that it is reasonable to ignore it altogether and adopt equations [3.2] and [3.3].

![Figure 3.3. Log-normal graph showing the various contributions made by the different production pathways of cosmogenic $^{10}$Be assuming a sea-level high-latitude production rate of 5.1 at/g/a. Fast and slow muogenic production becomes significant as denudation rates increase and, combined, can represent almost 25% of total production when denudation rates of 1 mm/a are occurring.](image-url)
3.3.4 Re-addressing the basin-wide denudation rate model:

It is useful to re-evaluate the validity of certain assumptions made in the Bierman and Steig (1996) model in light of the new formula for calculating denudation rates [3.6].

As has been discussed in section 3.3.2, mixing sediment from sub-basins of differing denudation rate, where the influence of the decay constant is variable, introduces error (equation [3.4]). Bierman and Steig (1996) calculate the significance of this error by considering a hypothetical basin composed of two sub-basins. One covering 5% of the total basin area and denuding at a rate which is one order of magnitude greater than the other sub-basin, which covers the remaining 95% of the total area. The sub-basin of higher denudation rate is also considered to be situated at greater elevation (i.e. a catchment headwall, or mountainous terrain within a larger catchment), hence it is given a production rate 3-fold greater than the more slowly denuding area which is considered to be at sea level\(^2\). The volumes of sediment denuded from each sub-basin are used to define the mixing ratio of their respective cosmogenic radionuclide concentrations. This cosmogenic radionuclide concentration mixture is then used in equation [3.3] to derive a denudation rate. The ‘assumed’ or ‘real’ denudation rate is calculated by spreading the combined denuded sediment volumes from each sub-basin back over the entire basin area. The degree of error introduced by mixing the sediments from these two sub-basins can then be

\(^2\) Bierman and Steig (1996) assume a \(^{10}\)Be production rate of 6.0 at/g/a and an \(^{26}\)Al spallogenic production of 36.0 at/g/a. The \(^{10}\)Be value has since been significantly refined (Stone, 2000). However, to allow comparison these values are adopted for the purpose of the model and it is assumed, though not made explicit by Bierman and Steig, that the modelled basin is at high latitude.
calculated by comparing the results of the cosmogenic radionuclide derived rate with the ‘assumed’ rate. This is essentially modelling the significance of the error in equation [3.4] using a range of values for $\varepsilon_1$ and $\varepsilon_2$. Here, this model is reconstructed but includes muogenic production according to equation [3.6] (Fig. 3.4) and the results compared with those of Bierman and Steig (Fig. 3.5). The significance of the error as it relates to typical San Bernardino Mountain elevations and latitude, with a refined value for $^{10}$Be production by spallation of 5.1 atoms/g/a, is also included.

**Figure 3.4** The cosmogenic radionuclide concentration, $N$, in sediment denuded from sub-basin 1 is mixed with that of sub-basin 2 in proportion to their respective sediment production ($\varepsilon$,%area). Sub-basin 2 has a 3-fold greater spallation production rate and corresponding muogenic increases, a ten-fold greater rate of denudation but only 5% of the total area. Values of $N_1$ and $N_2$ are calculated using equation [3.6].
Figure 3.5 The error introduced by mixing a. $^{10}\text{Be}$ and b. $^{26}\text{Al}$ in sediments from a basin denuding heterogeneously, where the decay constant has variable influence, is investigated. Numerical modelling of a catchment composed of two sub-basins, one comprising 5% of the total area is eroding at an order of magnitude faster than the other, comprising 95% of the area and whose rate is recorded on the x-axis. The log-normal plot shows that by including muons, the percentage difference in cosmogenic radionuclide derived denudation rate and the ‘assumed’ rate increases only slightly (red and blue lines) compared to the initial investigation into the effect by Beirman and Steig (1996) (green line) and that the error will be most severe when rates are low.

The introduction of muogenic production causes greater error, as would be expected considering the error results from the variable influence of the decay constant and that accounting for muogenic production at depth increases the time available for decay (section 3.2.2). The error is greater for $^{26}\text{Al}$ than $^{10}\text{Be}$ because the shorter half-life of $^{26}\text{Al}$ increases the significance of decay, but the error decreases for the
San Bernardino Mountain example due to muogenic production increasing less steeply with elevation and becoming less significant with respect to spallogenic production (Stone, 2000). However, if denudation rates over the majority of the basin are greater than $10^{-3}$ mm/a the error introduced for both cosmogenic radionuclides is small enough, relative to other errors associated with the technique, to ignore. Because uniform denudation rates across a basin are unlikely and the variable influence of the decay constant which would result, Bierman and Steig (1996) suggest ignoring decay altogether when rates are greater than several $10^{-3}$ mm/a. We can investigate whether this methodology remains sound when muogenic production is also being considered by comparing the percentage difference in denudation rates which would result from ignoring decay. The influence of ignoring the decay constant in equation [3.6] is considered for a sea level, high latitude drainage basin using the formula and spallation production rates advocated by Bierman and Steig (1996); the same production rates but including muons, as advocated by Granger et al. (2001), are used for a typical San Bernardino Mountain basin at a latitude of 34°N, an average elevation of 1500 m with the refined spallation production rate of 5.1 atoms/g/a (see footnote 2, page 40). Figure 3.6 shows the percentage increase in rates which would result erroneously if decay were ignored (no decay rate) in these scenarios. It is apparent that production of cosmogenic $^{10}$Be and $^{26}$Al by muons at depth makes the influence of decay more significant than for the same rates of denudation where only spallation is assumed. The error introduced into the spallation-only model for the no decay curve is small enough that it can be ignored providing denudation is greater than several $10^{-3}$ mm/a (Bierman and Steig, 1996). The inclusion of muogenic production requires
Figure 3.6 Bierman and Stieg (1996) suggest ignoring decay when deriving basin-wide rates. The difference in a. $^{10}$Be and b. $^{26}$Al derived denudation rate which results is plotted as a percentage on the y-axis and log denudation on the x-axis.

denudation rates on the order of $10^{-2}$ mm/a before the error associated with the no decay curve becomes small enough relative to other errors in the cosmogenic technique that it can be ignored. Providing rates are greater than $10^{-2}$ mm/a it should be possible to ignore the effects of decay and treat $^{10}$Be as a stable nuclide, in which case equation [3.6] can be simplified and rearranged to solve for denudation as in [3.7] below and no error is introduced by spatially heterogeneous denudation.
However, at such high rates this effect would likely also be negligible if decay was included (Fig. 3.5). For \(^{10}\)Be denudation rates of less than \(10^{-2}\) mm/a, decay should be accounted for and the assumption of whether or not spatial homogeneity exists to an extent where the variable influence of decay can be ignored may become problematic. These issues are even more pertinent to the use of \(^{26}\)Al.

\[
\varepsilon = \left[ \frac{P(o)}{N(o)} \cdot \frac{\Lambda}{\rho} \right] + \left[ \frac{YA_1}{N(o)} \cdot \frac{L_1}{\rho} \right] + \left[ \frac{YA_2}{N(o)} \cdot \frac{L_2}{\rho} \right] + \left[ \frac{B}{N(o)} \cdot \frac{L_3}{\rho} \right] \tag{3.7}
\]

The above discussion highlights the complex, interacting nature of the variables used to derive basin-wide denudation rates. Updating the theory of Bierman and Steig (1996) in the light of further understanding of cosmogenic radionuclide production by muogenic pathways does not invalidate the use of the basin-wide approach. It does, however, require that several further assumptions of their model be considered such as: 1. are denudation rates high enough that muogenic production must be accounted for; 2. are denudation rates high enough that decay can be ignored, and if not are they high enough that spatial heterogeneity of denudation within the basin of interest will introduce significant error? Significant error occurs for values between \(10^{-2}\) mm/a and \(10^{-3}\) mm/a, where muons are significant but so is their decay. As the influence of muons relative to spallation becomes less with elevation and lower latitudes, the effects described above, even over the
‘problematic’ range of denudation values, are unlikely to prevent valid application of the model.

3.4 Production rate calculations:

3.4.1 Measuring production rates:

With the development of accelerator mass spectrometry, and the ability to measure trace concentrations of cosmogenic radionuclides, came the need for accurate production rate measurements (Lal, 2000). Both theoretical and experimental methods have been used in order to obtain these values. Theoretical estimates of production rates have come from complex computer codes modelling the chance of cosmogenic radionuclide production based on particle energies and cross sections (Masarik and Reedy, 1995; Reedy, 2000). Artificial irradiation or exposure of synthetic targets has also been used to derive production rate values (Lal and Arnold, 1985; Heisinger et al., 2002a; 2002b; Nishiizumi et al., 1996). However, most applications of cosmogenic radionuclides have used the results from experimental data which attempts to record the concentration of cosmogenic radionuclide build up over a known time interval. Nishiizumi et al. (1989) provided an early estimate for production of in-situ cosmogenic $^{10}$Be and $^{26}$Al within quartz using the concentrations from glacially polished rocks in the Sierra Nevada to obtain respective values of 6.03 atoms/g/a and 36.8 atoms/g/a for combined spallation and muon production. Over following years, subsequent estimates for sea level, high latitude $^{10}$Be production ranged from 6.4 atoms/g/a (Brown et al., 1991) to 4.74 atoms/g/a (Clark et al., 1995) and this lack of resolution reduced confidence in the use of cosmogenic $^{10}$Be as a geochronometer (Gosse and Phillips, 2001). However,
Stone (1998b; 1999; 2000) recognised that the muogenic contribution to $^{10}\text{Be}$ at the earth’s surface had been overestimated and that by rescaling previous measurements, incorporating a refined value for the muogenic contribution, the discrepancies between the studies were resolved and they converged on an approximate sea level, high latitude value of $5.1\pm0.3$ atoms/g/a for spallation. A lack of resolution in production rates of $^{26}\text{Al}$ has been attributed, in part, to greater analytical error (Gosse and Phillips, 2001) but based on measured $^{26}\text{Al}/^{10}\text{Be}$ ratios a value of around six times that of $^{10}\text{Be}$ is typically assumed (Nishiizumi et al., 1989).

In this study sea level, high latitude $^{10}\text{Be}$ production is assumed to be $5.1$ at/g/a, based on Stone (2000), but a more conservative estimate of $10\%$ error rather than the $5\%$ assumed by Stone is employed due to remaining uncertainties in geomagnetic effects (section 3.4.2.1). Isolating the muogenic signature in bedrock in order to derive separate production rates from the spallogenic component has proved to be difficult (Stone et al., 1998a). For simplicity the sea level, high latitude muogenic production rate values based on modelled cross section data given in Granger et al. (2001) are used in this study.

3.4.2 Calculating basin-averaged production:

3.4.2.1 Variation in the geomagnetic field:

One of the fundamental assumptions stated in denudation rate models is that the production rate is uniform over time (Lal, 1991; Bierman and Steig, 1996; section 3.3.1). However, this is not entirely true for two reasons. Firstly, solar activity will influence galactic cosmic rays as they travel through the heliosphere. Evidence of
the influence heliospheric modulation has on cosmogenic radionuclide production is
difficult to measure but probably minimal and has generally been ignored (Gosse
and Phillips, 2001). Solar activity may also increase the proton flux of primary
radiation, although at energies so low that they will probably only slightly affect the
production rate at high (>60˚) latitudes (Masarik and Reedy, 1995).

Secondly, the flux and energy spectra of primary cosmic radiation reaching the
atmosphere is dependent on the cutoff-rigidity of the earth’s magnetic field. This
results in a latitudinal modulation of the cosmic radiation reaching the earth’s
surface which is catered for using scaling functions (Dunai, 2000; Lal, 1991; Stone,
2000; section 3.4.2.2). However, a further complication is that polar wander (or
secular variation) and magnetic palaeointensity variations drive latitudinal shifts of
the cutoff-rigidity of the earth’s magnetic field over time, causing cosmogenic
radionuclide production rates to vary temporally (Dunai, 2000; Gosse and Phillips,
2001). Much effort has been spent on quantifying the effects of geomagnetic
variation on cosmogenic radionuclide production rates (Clark et al., 1995; Dunai,
2000; 2001; Kubik et al., 1998; Masarik et al., 2001; Nishiizumi et al., 1989;
Snowball and Sandgren, 2002). However, predicting global correction factors is
hampered by bias introduced by the distribution of geomagnetic sample sites and by
the non-dipole component of the geomagnetic field (Dunai, 2001). In order to
account for changes in geomagnetic intensity and the effects of polar wander the
earth’s magnetic field has generally been considered to approximate a dipole,
however, this is an oversimplification and the actual geomagnetic field has a
significant non-dipole component which may contribute up to 25% of the total (Dunai, 2000; 2001; Guyodo and Valet, 1999).

Because of the spatial and temporal variation of production rates introduced by a changing geomagnetic field it is necessary to investigate this influence specifically for the latitude of the San Bernardino Mountains. Paleointensity data of McElhinny and Senanayake (1982) and Guyodo and Valet (1999) is used and the polar wander quantified using Ohno and Hamano (1992) in a methodology suggested by Gosse and Phillips (2001). For simplification, and due to a lack of agreement over the relative influence, any non-dipole effects are ignored. Furthermore, attenuation lengths will also tend to vary with geomagnetic effects as higher cutoff-rigidities allow only higher energy, more deeply penetrating, cosmic-ray-flux to pass through. However, this effect is considered secondary to production rate variation (Gosse and Phillips, 2001) and will not be considered here.

To quantify the effect of palaeointensity variations the present day geomagnetic latitude can be corrected to produce an ‘apparent’ latitude based on the severity of palaeogeomagnetic variation, such that, higher average past intensities will result in a higher apparent latitude (Gosse and Phillips, 2001). Production rates can then be scaled according to both the apparent and actual latitude and the two results compared. To achieve this the ratio of the palaeogeomagnetic dipole moment (integrated to the present) to its modern value (M/M₀) is used, as shown in equation [3.8] (Gosse and Phillips, 2001), where λ.geo is the apparent geomagnetic latitude and λ.pdgeo is the approximate latitude of the San Bernardino Mountains.
The palaeogeomagnetic dipole moment is taken from the McElhinny and Senanayake (1982) record for the last ten thousand years, which has a resolution of one thousand years. For comparison, more recent estimates of palaeomagnetic dipole moment over the last two to ten thousand years are used (Guyodo and Valet, 1999), which again have a resolution of one thousand years. McElhinny and Senanayake (1982) also provide a record for the last four thousand years with a five hundred year resolution. Results from all these data sets are plotted as a change in cosmogenic $^{10}$Be production rate over time and are shown for a range of altitudes in Figures 3.7 a, b and c.

To account for the changing magnetic field which would occur above a fixed point on the earth due to polar wander, the apparent geomagnetic latitude which would be obtained by integrating the changes over time at a specific point is derived using [3.9] below (Gosse and Phillips, 2001). Where $\lambda_P$ and $\Phi_P$ are coordinates of palaeopole positions from Ohno and Hamano (1992) and $\lambda_S$ and $\Phi_S$ are sample coordinates. For the latitude and longitude of the San Bernardino Mountains the $^{10}$Be production rate variation based on the above polar wander calculation [3.8] is shown in Figure 3.7 d.
\[
\cos \lambda_{geo} = \sin \lambda_s \sin \lambda_p + \cos \lambda_s \cos \lambda_p \cos(\Phi_p - \Phi_s)
\]

Both the intensity and polar wander effects on $^{10}$Be production appear appreciable and so in order to consider the influence on production rates in the San Bernardino Mountains their respective influence is combined for a range of altitudes and the results demonstrated for the different intensity records available in Figure 3.8. The results show that, depending on which record is used, production rates may be in error by ~20% in the San Bernardino Mountains at high elevations over particular timescales. The results also display the degree to which differing records will produce conflicting results. Until more dependable records of palaeogeomagnetic field intensity and polar wander are established it is surmised that any production rate correction based on the above data would be highly suspect. Accordingly, no corrections to production rates are applied in this study for geomagnetic effects. However, to compensate for the possible error introduced by an unquantifiable palaeogeomagnetic influence, the production rate error of ±5% assumed by Stone (2000) is replaced by a more conservative value of ±10%.

3.4.1.2 Atmospheric and latitudinal scaling:

The primary cosmic flux is composed mostly of charged particles which are deflected as they pass through the earth’s magnetic field. Thus, the geomagnetic
10ka record of palaeomagnetic intensity influence on Be-10 production based on McElhinny & Senanayke (1982)

Figure 3.7 (continued over)

0-4ka record of palaeomagnetic intensity influence on Be-10 production based on McElhinny & Senanayke (1982)
Figure 3.7 (continued) Graphs a, b, c and d show the influence of palaeogeomagnetic effects on the $^{10}$Be production rate over time at the latitude and for the elevation range of the San Bernardino Mountains. A sea level high latitude production rate of 5.1 at/g/a is assumed. The production rates are integrated over the time interval on the x-axis allowing straightforward interpretation as to the change of production over a specific averaging time. Curves in graphs a, b and c, over varying time periods, based paleointensity records are taken from McElhinny and Senanayake (1982) and Guyudo and Valet (1999). Graph d shows the effect of polar wander (or secular variation) on production rates based on the records cited in Ohno and Hamano (1992) and scaled by Gosse and Phillips (2001).
Figure 3.8 (continued over)
Figure 3.8 (continued) a, b, c and d. Series of graphs showing the effect of combined palaeogeomagnetic intensity variations and polar wander using the records shown in figure 3.7. Graphs a through d represent the production rate effect at elevations of 1, 2, 3 and 4 km, respectively. Depending on whether the record of McElhinny and Senanayake (1982) or that of Guyodo and Valet (1999) is used, the correction to production rates may range from less than 5 to nearly 20 percent. Also, by converting the record of McElhinny and Senanayake (1982) from one of a 500 year resolution to one of a 1000 year resolution, in order to splice it with the Guyodo and Valet record, error is introduced.
field acts as a filter allowing only high energy particles through at low latitudes but all energies through above 60° latitude, where the cutoff-rigidity of the magnetic field is lower than the low end of the energy spectra of cosmic radiation (Lal, 2000; Friedlander, 1989). For production of cosmogenic radionuclides in bedrock this means that latitudinal dependence must be taken into account. Furthermore, cosmic radiation is attenuated as it passes through the atmosphere and so production must be scaled for atmospheric pressure. Most applications using cosmogenic radionuclides have scaled their work using the functions presented in either Lal (1991) or Dunai (2001), although several other variations exist (Bierman et al., 2002).

Lal (1991) used neutron flux measurements taken from Geiger counters, photographic emulsions and cloud chambers to derive a third order polynomial fit with which sea level, high latitude production rates of cosmogenic $^{10}$Be and $^{26}$Al could be scaled for latitude and atmospheric pressure. This allowed production rates to be calculated for any location on the globe. In order to achieve this fit it was assumed the geomagnetic field approximated a dipole, thus removing any longitudinal variation, and the standard atmosphere model was used to convert atmospheric depth to elevation. Assuming a muogenic contribution to cosmogenic radionuclide production of ~15% Lal (1991, table 1) gives the coefficients (a, b, c and d) needed to apply equation [3.10] for any latitude, L, and altitude, y (km), to obtain in-situ $^{10}$Be or $^{26}$Al production at a point.
As noted above, however, it was found that a 15% muogenic contribution at the surface is an overestimate and that muogenic production increases less steeply with altitude and so must be scaled separately. In order to achieve this, table 2 of Lal (1991) should be used for the coefficients of equation [3.10], as it has no muogenic component, and the result divided by the nuclear disintegrations at sea level and high latitude (i.e. by 563.4). Stone (2000) recast table 1 of Lal (1991) as a family of fourth order polynomials for spallation only and this derivation can also be used to estimate spallogenic scaling. In order to scale the muogenic component for latitude and altitude, equation [3.11] is used (Stone 2000) where $M_{\lambda}(P)$ is the muogenic production rate scaling factor, $M_{\lambda,1013.25}$ is the scaling factor for latitude, given in Stone (2000, table 1), and $P$ is atmospheric pressure derived from the standard atmosphere model.

$$M_{\lambda}(P) = M_{\lambda,1013.25} \exp[(1013.25 - P)/242]$$

[3.11]

Dunai (2000) employs a slightly modified neutron flux data set to Lal (1991) and uses geomagnetic field inclination when scaling for latitude in order to account for non-dipolar geomagnetic effects. Furthermore, both the absorption mean free path and production rate are scaled for altitude and latitude. In this study, the corrections suggested by Lal (1991) are selected because of disagreements concerning Dunai’s
use of neutron flux records and the relative merits of using field inclination to order neutron intensity data (Desilets et al., 2001). However, in order to consider the differences between the two methods the rate of spallation of cosmogenic $^{10}\text{Be}$ derived using both methods is compared below.

Scaling production rates according to atmospheric pressure and latitude is relatively straightforward for site-specific samples but for the basin-wide approach an average must be considered that is representative of the mean atmospheric shielding and latitude of the basin. In order to achieve this digital elevation models were used. Figure 3.9 shows the exponential relationship between the rate of production by spallation and elevation at the latitude of the San Bernardino Mountains. Muogenic production is ignored in this graph for simplicity but it should be noted that it is

![Graph showing the exponential decrease of production rate with increasing atmospheric depth (decreasing altitude) for 34ºN latitude. Based on the scaling of Stone (2000, table 1) which has been recast from the scaling of Lal (1991, table 2) assuming a sea level high latitude production rate of 5.1at/g/a.](image-url)

**Figure 3.9** Graph showing the exponential decrease of production rate with increasing atmospheric depth (decreasing altitude) for 34ºN latitude. Based on the scaling of Stone (2000, table 1) which has been recast from the scaling of Lal (1991, table 2) assuming a sea level high latitude production rate of 5.1at/g/a.
attenuated less, hence, increases less steeply with elevation and should be scaled separately (Gosse and Phillips, 2001). Because of the non-linear relationship between production and atmospheric depth, the mean elevation of the basin will not correspond to the mean production rate of the basin, it is instead a function of hypsometry. Therefore, the production rate for each digital elevation model cell can be calculated according to altitude and latitude and then integrated across the basin under consideration to obtain the mean rate of production. This is done using ArcView 3.2 software and a detailed description of this procedure is given in Appendix 2. Figure 3.10 shows a digital elevation model of a basin with >2 km relief in the San Bernardino Mountains and depicts the spatial variability in $^{10}$Be production rates for spallation, slow and fast muons according to Lal (1991) and Granger et al. (2001). Of note is the strong altitudinal dependence of all production pathways. The difference between rates of spallation calculated using Lal (1991) and Dunai (2000) is shown in Figure 3.11. As can be seen the maximum difference in production rates predicted by the two methods is ±5%.

3.4.3 Shielding issues:

3.4.3.1 Slope:

If samples are collected from a sloping surface part of the sky will be blocked, preventing a complete 360° hemispherical exposure to incoming cosmic radiation and reducing the rate of production and attenuation length of in-situ-produced cosmogenic radionuclides. Similarly, if distant topography blocks out part of the horizon, less cosmic radiation will strike a particular sample site (Dunne et al., 1999). These are termed ‘slope-angle’ and ‘topographic’ shielding, respectively.
Figure 3.10 Basin-wide latitudinal and altitudinal scaling for production of in-situ-produced $^{10}$Be by a. spallation b. slow muons and c. fast muons. Basin size is 19km$^2$.
Figure 3.11 Digital elevation model of a basin in the San Bernardino Mountains illustrating the difference in $^{10}$Be production rates between the atmospheric and latitudinal scaling of Lal (1991) and Dunai (2000). Sea level, high latitude production rate is 5.1 atoms/g/a. Note, Lal’s scaling is greater at low elevations (~1km in this basin) and Dunai’s greater at high elevations (~3km in this basin) but the difference does not exceed 5%. As this basin displays almost the full range of elevation present in the San Bernardino Mountains, and the effect of latitudinal difference is comparatively small, the difference between selecting either Lal or Dunai’s scaling techniques in these mountains is within the error associated with the $^{10}$Be production rate.

The effects of shielding can be catered for in a relatively straightforward manner for site-specific bedrock samples by calculating the degree of open sky available to a bedrock outcrop and measuring the slope of the portion of the surface removed (Dunne et al., 1999; Gosse and Phillips, 2001; Masarik et al., 2000). Less discussion has been presented on the effects of sample shielding for basin-averaged production rates. How these effects have been dealt with in this study is considered below.

Cosmic radiation is attenuated as it passes through the atmosphere. This means that the flux will have the least depth of atmosphere through which to pass when the angle of incoming radiation (incident angle) is directly overhead, and the greatest
depth as the incident angle tends towards the horizon. Because of this, incoming radiation does not strike the earth’s surface isotropically but the majority of radiation will approach a surface from directly overhead (i.e. perpendicular to a surface of 0° slope). Until a slope faces away from the overhead significantly (the most extreme case being a vertical, or overhanging cliff face) production rates will be only slightly reduced and for the same reasons surrounding topography must be significantly above the horizon to have a significant effect. Figure 3.12 shows the relationship for production at the surface due to slope angle shielding based on the formula given in Dunne et al. (1999, equation 18) and given below [3.12], where $S$ is the scaling factor and $\alpha$ is the slope angle.

$$S = 1 - 3.6 \times 10^{-6} \alpha^{2.64}$$

[3.12]

**Figure 3.12** The scaling factor for cosmogenic radionuclide (CRN) production rates based on slope angle using the formula of Dunne et al., (1999) shown in equation [3.12].
One approach which has been suggested to apply slope angle shielding to whole basin production rates is to use digital elevation models to calculate the shielding for every digital elevation model cell within the basin of interest. Integrating the results across the basin produces a production rate scaling factor for mean slope angle shielding\(^3\). However, the flux of cosmic radiation which impinges upon a basin’s surface, hence the cosmogenic radionuclide reservoir of a basin, is not influenced by internal slopes (or other obstructions) such that once the radiation has passed below the drainage divide internal slope angles are irrelevant (Riebe, 2002, pers. comm.). In this respect the important factor is the angle away from the horizontal at which the basin lies.

One way to visualise this is the angle, from the horizontal, at which a hypothetical flat, but inclined, surface laid across the drainage divides would sit as represented by a plane going through the basins river outlet point and the point of highest elevation. Figure 3.13 illustrates this. In the case of a large lowland basin the surface would be only slightly tilted up from the horizontal, whereas, the drainage divides of a steep mountainous catchment would display a greater difference in elevation and so the hypothetical surface would be tilted more and the basin more shielded. The extreme case, again, is a cliff face where the drainage divides, for the purpose of this argument, are the top and bottom of the cliff and the hypothetical surface lies perpendicular to the horizontal. Correcting production rates using this latter approach is more consistent with the concept of the basin acting as a cosmogenic

\(^3\) See http://depts.washington.edu/cosmolab/Pby%20GIS.html
The flux of cosmogenic radiation which impinges on a basin is a function of the overall slope of that basin, here represented by the basin long-axis angle $\alpha$. The basin is shielded from incident radiation which passes through the atmosphere between the horizon and the angle $\alpha$. For large lowland basins as represented by Figure A, this slope angle, $\alpha$, is small and the correction to production rates minimal. For small high relief catchments found in mountainous terrain where $\alpha$ is greater, as illustrated by Figure B, the correction becomes more significant.

Figure 3.13  The flux of cosmogenic radiation which impinges on a basin is a function of the overall slope of that basin, here represented by the basin long-axis angle $\alpha$. The basin is shielded from incident radiation which passes through the atmosphere between the horizon and the angle $\alpha$. For large lowland basins as represented by Figure A, this slope angle, $\alpha$, is small and the correction to production rates minimal. For small high relief catchments found in mountainous terrain where $\alpha$ is greater, as illustrated by Figure B, the correction becomes more significant.

In order to measure the affect on production rates this form of shielding has, the maximum and minimum elevations of a basin obtained from digital elevation model data are used to define the maximum and minimum drainage divide elevations. The radionuclide reservoir in the methodology considered by Bierman and Steig (1996).
distance between these points is then used to define the angle from the horizontal at which the hypothetical surface would sit (or the angle of the basin long-axis) and equation [3.12] is used to determine the scaling factor which will be applied to the production rate to account for the fact that the basin is tilted and not exposed to 360° of incoming radiation. This calculation is an approximation only as the shape of the basin is unlikely ever to be as idealised as discussed; however, the corrections which apply to production rates because of this shielding are generally small with the largest in this study being 6% and the majority being <1%. Furthermore, shielding by distant topography in the San Bernardino Mountains is negligible and not considered, although basins facing each other on opposing valley sides may introduce a small correction factor it does not exceed 1% and so is not included.

3.4.3.2 Snow shielding:

A further method by which mineral targets may be shielded from incoming cosmic radiation is by snow cover (Gosse and Phillips, 2001). Above ~1250 m elevation the San Bernardino Mountains presently receive around six months of snowfall per year, the amount of which varies according to elevation (Minnich, 1986; Minnich, 1989). At around 3000 m elevation, there is <4 m of mean annual snowfall and below 1250 m snow is rare (Minnich, 1986). While palaeosnowfall cannot be quantified, greater annual snowfall would be expected during periods throughout the Holocene when there is evidence of several small glaciers at the highest elevations in the San Bernardino Mountains (Ingle and Moran, 1958; Sharp et al., 1959; Owen et al., 2003). Therefore, although modern day estimates of snowfall are used, they should be considered minima.
The effect of snow shielding is quantified using the formulation given by equation [3.13], where \( \rho_{\text{snow}} \) is the density of snow, taken here to be 0.24 g/cm\(^3\) (Gosse and Phillips, 2001); \( x_{\text{snow}} \) is the depth of snow fall occurring per month and \( \Lambda \) is the attenuation length. The measurements of shielding per month are then integrated over the year, depending on how many months of the year snowfall is estimated to occur, to give \( P_{\text{base}} \), the production rate at the base of the snow layer.

\[
P_{\text{base}} = \frac{1}{12} \sum_{t=1}^{12} P_{\text{surface}} e^{\left(-\left(\frac{\rho_{\text{snow}} x_{\text{snow}}}{\Lambda}\right)\right)}
\]  

[3.13]

Snow shielding at greatest elevations in the San Bernardino Mountains will result in \(<2\%\) and \(<1\%\) error in production rates for slow and fast muons, respectively, so this correction is ignored in light of the more significant sources of error of the cosmogenic radionuclide technique. The error to rates of spallation is greater due to the shorter attenuation length of the neutron flux in cosmic rays and this effect is estimated for a range of annual snowfall depths assuming 6 months snow cover (Minnich, 1986) (Fig. 3.14 a). Based on Minnich (1986; 1989), a 4 m annual snow depth is assumed at 3000 m which declines linearly to zero at 1200 m, below which no snow falls. The scaling factor for spallation is derived for this range of elevations and is applied to basin mean elevations (Fig. 3.14 b). Basins in the northern regions of the mountains are excluded from any corrections as they are in the lee of the rain shadow created by the elevated southern San Bernardino Mountains.
**Figure 3.14 a.** Graph showing the percentage decrease in production rate for the range of annual snowfalls in the San Bernardino Mountains, assuming 6 months snow cover based on the records Minnich (1986) recorded over the last several decades. **b.** The correction applied to the cosmogenic radionuclide production rates in the San Bernardino Mountains. The ‘knee’ at 1200m is due to the rarity of snowfall recorded below this altitude. These scaling values should be considered minima as palaeosnowfall in the San Bernardino Mountains was likely higher during the Holocene, based on evidence of glaciation (Ingle and Moran, 1958, Sharp *et al*., 1959, Owen *et al*., 2003).

### 3.5 Averaging time:

An issue central to any measurement of rate is the length of time over which the measurement takes place, or the averaging time of the measurement. The averaging
time of the denudation rates measured by cosmogenic radionuclide analysis is a function of how long a target mineral is exposed to cosmic radiation, that is to say it is the length of time over which cosmogenic radionuclides have been accumulating. This time span will depend on two variables, the depth at which cosmogenic radionuclides are produced, as defined by the absorption coefficient of the incident radiation, and the rate at which those cosmogenic radionuclides are brought to the surface, defined by the denudation rate.

The absorption coefficient, described in section 3.2.1, is a function of the attenuation length and the density of that matter \( \rho/\Lambda \) giving it units of \( \text{cm}^{-1} \). A more intuitive interpretation is to consider the inverse of the absorption coefficient (i.e. \( \Lambda/\rho \)) as a coefficient of penetration, measured in units of cm (equation [3.3], section 3.3.1). Thus, the penetrative ability of cosmic radiation will be greatest when the flux of cosmic radiation is ‘hard’ giving it a greater attenuation length (i.e. it is composed of radiation from the high end of the energy spectrum) and when the density of the matter being penetrated is low. The fast and slow muogenic, and nucleogenic components of cosmic radiation have decreasing attenuation length, respectively, resulting in a range of depths over which they are produced and making their influence on the averaging time a weighted function of these depths. This is explained in more detail below.

Considering the simplest case of spallation alone (equation [3.2]) with an attenuation length of between \( \sim140-170 \text{ g/cm}^2 \) in a granite body with a density of \( \sim2.6 \text{ g/cm}^3 \) (Gosse and Phillips, 2001); the coefficient of penetration will be \( \sim60 \text{ cm} \). This is the
depth of rock removed in the denudation rate model and the time required to remove this depth is the averaging time of the technique. Subsequently, it is relatively easy to envisage how a more rapid rate of denudation will incorporate a shorter averaging time if the coefficient of penetration remains the same. This will be true if lithologies do not vary significantly and the attenuation length remains constant. As the attenuation length is dependent on the cutoff-rigidity of radiation it is a function of magnetic field and atmospheric depth (Dunai, 2000; Muzikar et al., 2000). However, the variation this produces is small and has generally been ignored in cosmogenic radionuclide applications, although there remains some uncertainty as to the value to use for attenuation lengths. In this study the value of $160\pm10$ g/cm$^2$ is adopted to remain consistent with the model formulation by Granger et al. (2001) and it is assumed to be constant with latitude and elevation. This gives a penetration coefficient of 61.5 cm for spallation. In order to include muogenic production and derive averaging times when applying equation [3.7] the longer attenuation lengths but lower production rates of muons must be considered. Granger et al. (2001) give attenuation lengths for two different slow muon energies of 738.6 g/cm$^2$ and 2688 g/cm$^2$ and a fast muon attenuation length of 4360 g/cm$^2$, resulting in penetration coefficients of 284 cm, 1034 cm and 1676 cm, respectively, assuming attenuation is occurring in bedrock with a density of 2.6 g/cm$^3$. In order to obtain an averaging time which reflects the production of muons at depth a weighted mean of all the attenuation lengths, weighted by relevant production rate, was calculated and from this an averaging time obtained. As muogenic production scales differently with elevation than spallation, and the influence of muons vary with denudation rate, the relative influence of each production pathway on the averaging time also varies. To
appreciate the difference that including muons has on the averaging time, Figure 3.15 displays averaging times for production rates both with and without a muogenic component.

Figure 3.15 Log-log plot of the relationship between the averaging time of the cosmogenic radionuclide analysis and the rate of denudation when spallation and muogenic production is considered (thin black line) as in equation [3.7] and when only spallation is assumed (thick grey line) as in equation [3.3]. Averaging times increase more than four-fold when muogenic production is considered for denudation rates of 1 mm/a. Note also that decay is ignored, as discussed in section 3.3.4.

The model of denudation requires the assumption that rates have been consistent enough over time that a secular equilibrium between the flux of cosmogenic radionuclides being produced in, and leaving a basin, has been achieved. The length of time that this secular equilibrium has existed for is three to five times the averaging time; as the removal of 2-3 m of bedrock (i.e. three to five times ~60 cm) is required to eradicate the cosmogenic radionuclide signature of a previous rate (Bierman and Steig, 1996; Lal, 1991). However, Lal (1991) notes that secular equilibrium does not necessarily require perfectly uniform denudation over this time,
and such a situation is unlikely. Because the contribution by muogenic production will be sensitive to rates of denudation, which may vary slightly through time, the weighted mean attenuation length and hence the averaging time will also vary. As such the averaging times should be considered approximations.

3.6 Cosmogenic radionuclide sampling and measurement:

In order to interpret a basin-wide denudation rate, cosmogenic radionuclide concentrations in alluvial sediments leaving the basin must be measured. This involves the collection of alluvial material, the separation of clean quartz from the sample, the extraction of the cosmogenic $^{10}$Be and $^{26}$Al from the quartz and then measurement of these concentrations by accelerator mass spectrometry. In the following section each of these will be discussed in terms of the target preparation procedures used in the University of Edinburgh Cosmogenic Laboratories and accelerator mass spectrometry analysis in the Department of Nuclear Physics at the Australian National University. Where appropriate, techniques used in other facilities will be mentioned. The protocols used to prepare the San Bernardino Mountains sample targets have evolved over time as methods increasing the processing efficiency and sample purity have been tested and the discussion below represents the present culmination of that development. For more detailed protocols of the sample collection and preparation see Appendix 1.

3.6.1 Sample collection:

The collection of alluvial samples for basin-wide denudation rate applications requires careful consideration of the geomorphological system being sampled.
Glacial activity, deep-seated landsliding, anthropogenic activity and heterogeneous lithology may all invalidate the use of the model by breaking the assumptions that production rates across a basin have been temporally uniform (palaeogeomagnetic variation accepted) and that sediments are thoroughly mixed and representative of the entire area upstream. If the application is basin-wide denudation rates, the sediments should be collected from the active channel whenever possible to reduce the possibility of sampling stored sediments. For small mountainous streams this may be straightforward but in the case of larger lowland streams with more appreciable channel width, or arid ephemeral environments where the channel may be hard to define, an amalgamation of sediment from across the channel or wash may be appropriate (Schaller et al., 2000; Clapp et al., 2002). In summary, an appreciation of the environment sampled, and the degree to which any of the above could introduce bias, is required.

The grain size required from alluvial samples is a trade off between avoiding fine grained sediments, that have had aeolian transport and may have come from out-with the basin, and too large a grainsize that that may increase the likelihood of including non-quartz inclusions or composite grains. A typical range is somewhere between 125 μm to 1000 μm (Clapp et al., 2001). Whether or not there is a bias of cosmogenic radionuclide concentrations due to grainsize is controversial and will be considered further in Chapter 4.

The volume of sediment which must be collected should be based on estimates of the quartz content of the alluvium and the approximate denudation rate expected.
Sediment from more rapidly denuding environments will contain less cosmogenic radionuclides per gram and in order to measure this concentration above the background, and to reduce error, more quartz will be required than for slowly denuding terrains. Between 20 and 60 g of clean quartz is desirable from each sample depending on the caveats discussed above but it should also be noted that some quartz will be lost during the cleaning process.

3.6.2 Accelerator mass spectrometry target production:

3.6.2.1 Quartz extraction:

The first stage of sample processing is to obtain pure quartz separates from the sieved alluvial sample. The following describes the procedures developed at the University of Edinburgh Cosmogenic Laboratories. The protocols presented here may deviate slightly from standard ones which have been developed at the University of Edinburgh due to necessities borne from using the basin-wide approach in a rapidly denuding area (Chapters 6 and 7). However, both variations derive principally from the techniques used at the University of Vermont (Bierman et al., 2002) where the protocols have been adapted from the techniques of others and are similar, in principle, to the techniques employed at most other laboratories.

While olivine has been used in cosmogenic $^{10}$Be and $^{26}$Al denudation rate applications (Nishiizumi et al., 1989), quartz is typically the mineral of choice for many reasons. It is abundant and resistant to depletion in most environments, it contains low $^{27}$Al abundances and can be effectively cleaned of atmospheric $^{10}$Be (often termed ‘meteoric’ or ‘garden variety’) (Bierman, 1994; Lal and Arnold, 1985;
Liese, 2002). The most time consuming and costly stage in the preparation of clean quartz is acid etching. If the quartz concentration of a sample can be increased with other separation techniques initially, etching becomes more efficient. Removing non-quartz minerals can be achieved using a variety of methods. Mafic minerals may be removed using magnetic separation techniques. These range from brushing the sample with hand magnets to Franz electromagnetic separation. Density separation of minerals in heavy liquids using lithium polytungstate solution (LST) allows minerals of varying density to be segregated in a funnel by floating or sinking them. However, such techniques are efficient only if required sample sizes are small or if the sample is rich in mafic minerals. If quartz concentrations are low the amount of sample which must be processed may be inappropriate for the use of electromagnetic or heavy liquid separation and if grain sizes are small (<250 μm) density separations will be problematic as the weight to surface area ratio of small grains prevents realistic mineral segregation (Barrows, 2002, pers. comm.).

For large samples, one or more ‘junk’ etches may significantly increase quartz concentrations. For granitic samples collected from the San Bernardino Mountains, junk etching typically increased quartz yields from ~20% to >30%. Junk etching involves placing several hundred grams of sample in around 1 l of diluted HF and leaving it over several days on a shaker table or ultrasonic bath. It is usually necessary to acid clean the sample in HCl to clean Al and Fe coatings from grains (Bierman \textit{et al.}, 2002) and remove any carbonates so this stage should preclude the junk etch to reduce the amount of sample drying time. Around 25 g of sample is then given at least three 24 hour etches in a 2.5 l mixture of 1% HF and 2% HNO₃.
which both removes non-quartz minerals and etches the surface of the quartz grains, removing any adsorbed atmospheric $^{10}\text{Be}$. The $\text{HNO}_3$ prevents build up of a silica gel in the acid solutions but many laboratories successfully use only HF (Stone, 2003, pers. comm.). Some laboratories use pyrophosphoric acid treatments to etch samples which may allow a reduction in processing time (Riebe et al., 2000). If density separations have been performed at an earlier stage the sample needs only one more etch; if not, density separations will remove any undissolved mafic minerals before this last etch, which should result in a pure quartz separate. As the surface of the quartz grains is removed during etching, and smaller grain sizes have larger available surface area, the quartz lost during etching will be a reflection of the grain size selected. For San Bernardino Mountain granites with ~25% quartz concentrations it was often necessary to etch >250 g of 125-250 μm size fraction to achieve pure quartz masses of <50 g, implying a significant proportion of the quartz was lost during etching. A larger grain size would thus have been preferable but could not be used in this situation because of the likelihood of inclusions within larger quartz grains. It is advisable at this stage to perform Inductively Coupled Plasma, or Atomic Absorption mass-spectrometry, on an assay of the sample. If Al concentrations are greater than ~200 ppm, further etching should be performed.

### 3.6.2.2 Cosmogenic radionuclide extraction:

Once a sufficient amount of clean quartz separate has been obtained (typically between 20 to 60g) the targets for accelerator mass spectrometry analysis can be produced in the form of oxides of Be and Al. The cleaned quartz sample is dissolved over heat in concentrated HF. A sample blank is also introduced at this
stage to measure any sample loading which may occur. Prior to dissolution, a Be carrier must be added to all the samples and the blank. Accelerator mass spectrometry measures isotope concentrations by measuring the ratio between two isotopes. $^9\text{Be}$ rarely occurs in samples (Bierman et al., 2002) and so a known quantity (the carrier) is added to produce a $^{10}\text{Be}/^{9}\text{Be}$ ratio. The amount of carrier added is a trade-off between errors and target handling. More carrier will make the physical handling of the target at later stages easier and allow longer accelerator mass spectrometry beam time measurement, lowering this source of error. However, more carrier will also lower the isotopic ratio of the target and can make it difficult to measure the sample above the ratio of the sample blank, thus increasing this source of error. Purchased Spectrosol $^9\text{Be}$ 1000 mg/l standard solution was the carrier used for San Bernardino Mountains samples, which has a background $^{10}\text{Be}/^{9}\text{Be}$ ratio on the order of $10^{-14}$. For some rapidly eroding sites where 350 μg of carrier had been added, the $^{10}\text{Be}/^{9}\text{Be}$ ratio approached this value and it became difficult to observe the sample ratios above those of the blank. In such cases it is advisable to increase the amount of quartz dissolved, add smaller carrier amounts and ideally make a $^9\text{Be}$ carrier using deep mined Beryl which should have a $^{10}\text{Be}/^{9}\text{Be}$ ratio on the order of or less than $10^{-15}$. Carriers as small as 150 μg were used in San Bernardino Mountain samples with no significant reductions in the accelerator mass spectrometers beam current. However, carrier additions smaller than this would be impractical to handle and 200-250 μg carriers would be more appropriate in all but the most rapidly denuding (hence low concentration) samples. Co-precipitation of Be with silver nitrate has the potential to add even small carrier amounts while retaining the manageability of the sample (Stone et al., in press);
however, this procedure was not performed on the San Bernardino Mountains samples. Al carriers may also be required if the Al assay has revealed samples contain <3000 µg Al (Bierman et al., 2002). This is usually not essential as samples will likely already contain appreciable $^{27}$Al concentrations, revealed during the Al assay. However, using Al carrier allows an Al sample blank to be introduced.

Hotplate temperatures during sample dissolution should not exceed 180°C to prevent evaporation of Al complexes. Samples <30 g may dissolve in a few days but larger samples can take several weeks. After dissolution a second Al aliquot should be taken for Inductively Coupled Plasma, or Atomic Absorption mass-spectrometry analysis if $^{26}$Al is to be measured. This provides the $^{27}$Al mass for the accelerator mass spectrometry $^{26}$Al/$^{27}$Al ratio measurement.

Samples are then fumed several times by addition of HClO$_4$ on a hotplate. This removes unwanted fluorides by converting them to perchlorates and evaporating them (Bierman et al., 2002). After the samples have been fumed they are taken up in solution and purified by passing them through anion exchange columns. Anions of Fe in the sample will preferentially be adsorbed onto the column resin in the presence of strong HCl, whereas the Al and Be cations in the sample will be eluted. Any Ti in the eluted sample can be precipitated out at ~pH 4; >200 µg of Ti per target can significantly reduce beam currents (Fifield, 2002, pers. comm.). However, some have reported loss of Be by precipitation at this pH also (Child et al., 2000). The purified sample is then passed through cation exchange columns where the Be and Al are separated by being adsorbed and eluted under differing HCl
concentrations. HF and H$_2$SO$_4$ have also been used for column chromatography, although care must be taken as sulfuric acid may facilitate complex compounds with Be (Child et al., 2000; Bierman et al., 2002; Barrows, 2002, pers. comm.). The resulting Be and Al elutants are precipitated at pH ~8 and ~9 respectively and left overnight so that B will go back into solution and can be decanted in the supernatent. As an isobar of $^{10}$Be, B can cause difficulties during accelerator mass spectrometry measurement (section 3.6.3) and levels should be reduced as much as possible during target preparation. Foremost, this involves removal of any borosilicate glassware from the post-etching stages of the processing to prevent contamination. Several water rinses and re-precipitations should be performed as this helps remove further impurities and forms solid ‘plugs’ of sample during the final dry down, making handling easier. A final HClO$_4$ fume of the Be sample fraction before the final dry down will also help reduce B levels (Bierman et al., 2002). Precipitated BeOH and AlOH gels are dried down at low temperatures. The ‘plugs’ that form will typically be <0.5 mm in diameter, depending on the amount of carrier added and can range in colour from white to dark brown. These are fired in quartz vials to form Be and Al oxides which are then mixed with Nb or Cu (for BeO) and Ag (for AlO) powder and pressed into sample cathodes for loading into an accelerator mass spectrometry wheel. In the case of the San Bernardino Mountain samples this was undertaken at Australian National University using a ratio of around 1 part sample to 8 parts Nb for BeO.
3.6.3 Accelerator mass spectrometry analysis:

Accelerator mass spectrometry has the ability to measure trace concentrations of elements by accelerating their ions, deflecting their trajectories using powerful magnets and recording their relative concentrations. Figure 3.16 displays a typical accelerator mass spectrometry schematic and the discussion below will consider the function of this apparatus when measuring cosmogenic radionuclides. Much of the discussion is based on the measurement of San Bernardino Mountain samples at the accelerator mass spectrometry facility at the Australian National University and it should be noted that the workings of this facility may differ slightly from others. Furthermore, the analytical precision of Be measurement is better than that of Al and it is Be which is used to derive denudation rates in later chapters so most of the discussion will centre around it.

Packed targets are placed in front of a Cs ‘sputter’ source, the ‘negative ion source’ in Figure 3.16. The beam of ionised Cs is directed towards the target cathode containing the BeO or AlO which sputters ions from the cathode and they are then pre-accelerated. In the case of Be, $^{10}$Be$^{16}$O$^-$ and $^9$Be$^{17}$O$^-$ are injected into the accelerator. The beam of negative ions is then accelerated towards the high voltage terminal which is given a large positive charge state by electrostatic tandem pelletron belts. Upon reaching the terminal the $^{10}$Be$^{16}$O$^-$ and $^9$Be$^{17}$O$^-$ are stripped of their ions by a gas or foil stripper or both (Fifield, 1999). Large accelerators, such as the National Electrostatics Corporation 14UD Pelletron employed at the Australian National University, are able to strip Be of enough electrons to produce $^{10}$Be$^{+3}$, while smaller accelerators may only obtain a +2 charge state.
Figure 3.16 Schematic of an accelerator mass spectrometer showing the principal components discussed in the text. Adapted from Fifield (1999).
The positive $^{10}\text{Be}$ ions are then repelled away from the terminal, further accelerating them till they reach the analysing magnet. This deflects ions of varying weights to different detectors. An advantage of high voltage accelerator mass spectrometers is that it they are able to produce $^{10}\text{Be}^{3+}$ with the same magnetic rigidity, and hence the same trajectory through the analyser magnet as $^{17}\text{O}^{5+}$, which is produced by the stripping of $^9\text{Be}^{17}\text{O}^-$. The $^{17}\text{O}$ count acts as a measure of the amount of $^9\text{Be}$ and can be recorded simply by a Faraday cup. Without this advantage, a switching between $^{10}\text{Be}$ and $^9\text{Be}$ sources is required which can affect the stability of the terminal voltage (Fifield, 2002, pers. comm.). $^{10}\text{Be}$ is measured at the Australian National University by an ionisation chamber. Here, a gas filled chamber slows the $^{10}\text{Be}$ ion producing electrons. These electrons drift toward anodes along the length of the chamber where the charge is recorded. Because $^{10}\text{Be}$ and $^{10}\text{B}$ lose energy at different rates they each have identifiable signatures. These signatures are recorded by the anodes along the detector which allows discrimination between the isobars giving the $^{10}\text{Be}$ count. Each sample is run, typically, for a period of several minutes and these runs are usually repeated two or three times if there is enough sample left to sputter. Throughout the analysis, standards (NIST standards in the case of the Australian National University accelerator mass spectrometer) are run with every ten or so samples in order to identify any drift and to correct the $^{10}\text{Be}/^{17}\text{O}$, and by proxy, the $^{10}\text{Be}/^9\text{Be}$ ratios recorded.
3.6.4 Cosmogenic nuclide concentration calculation:

Accelerator mass spectrometry returns a $^{10}\text{Be}/^{9}\text{Be}$ ratio which then must be converted to a measure of $^{10}\text{Be}$ concentration in order to derive denudation rates (section 3.3). By measuring the $^{10}\text{Be}/^{9}\text{Be}$ ratio present in a sample, and knowing the mass of the quartz separate and amount of $^{9}\text{Be}$ carrier which has been added (section 3.6.2.2) it is possible to calculate the $^{10}\text{Be}$ concentration in atoms per gram of sample. However, in order to account for any non-cosmogenic $^{10}\text{Be}$, which may have contaminated the sample during the post-etching stages of the sample preparation, a sample blank is used (section 3.6.2.2). The number of $^{10}\text{Be}$ isotopes present in the sample blank, measured by accelerator mass spectrometry, is subtracted from each sample prepared alongside the blank to account for any sample loading. Sample blank $^{10}\text{Be}/^{9}\text{Be}$ ratios of $10^{-16}$ atoms/g are possible but $10^{-15}$ atoms/g, or $10^{-14}$ atoms/g are more common.

Equation [3.14] is used to derive radionuclide concentrations where: $\text{avo}$ is a constant termed Avogadro’s Number, $6.022 \times 10^{23}$; $A$ is the atomic number of the element being measured, 9.013 in the case of Be; $^{10}\text{Be}/^{9}\text{Be}$ and $^{10}\text{Be}_{\text{blank}}/^{9}\text{Be}_{\text{blank}}$ are the ratios measured by accelerator mass spectrometry for the sample and the blank, respectively; $c$ is the mass of carrier added (g); and $m$ is the mass of pure quartz separate which was dissolved (g).

$$ N = \left( \frac{^{10}\text{Be}/^{9}\text{Be}}{c \cdot \text{avo} / A} \right) m - \left( \frac{^{10}\text{Be}_{\text{blank}}/^{9}\text{Be}_{\text{blank}}}{c \cdot \text{avo} / A} \right) $$

[3.14]
3.7 Summary:

The above has detailed the workings of the basin-wide denudation rate model, the assumptions it makes and the variables in incorporates. With the advent of a model of denudation which incorporates muogenic production some of these assumptions need to be revalidated and so have been empirically tested here. They are shown to be consistent with a basin-wide denudation rate model incorporating muogenic production. Secondly, a discussion of the scaling and corrections which need to be applied to make accurate estimates of cosmogenic radionuclide production rates has been given and the specific calculations used in the measurement of rates in the San Bernardino Mountains presented. The concept of the averaging time of cosmogenic radionuclide analysis has received separate attention not only because it was required to be re-addressed in the light of muon production but because it is particularly relevant to discussions in Chapters 6 and 7. Lastly, issues of sample collection, processing, analysis and measurement have been discussed and the techniques specific to sampling in rapidly denuding terrain such as the San Bernardino Mountains have received particular attention. While many of the assumptions incorporated by basin-wide cosmogenic radionuclide analysis can be addressed by numerical modelling, there are some issues which require testing in the field. In the following chapter this theme is developed by focusing on sediment mixing and processes of denudation operating in the San Bernardino Mountains.
4. Sediment mixing in the San Bernardino Mountains

4.1 Introduction:

Since initial testing, the application of cosmogenic radionuclide analysis using the basin-wide approach has prospered, with the technique being employed in a range of environments from lowland desert settings (Bierman and Caffee, 2001; Clapp et al., 2001; 2002), to tropical mountainous regions (Hewawasam et al., 2003; Brown et al., 1995; 1998). However, with the exception of Vance et al., (2003) there has been a paucity of studies in tectonically active regions undergoing rapid denudation and still less testing of the technique in these areas. This chapter seeks to redress this imbalance by questioning the validity of assumptions of sediment mixing, vital to ensuring that the basin-wide approach can be employed in high relief tectonically active orogens. The study will focus on regions of rapid denudation within the San Bernardino Mountains, southern California, and consider the following questions. Firstly, is sediment mixing in channel reaches sufficient enough that the cosmogenic radionuclide concentrations obtained from alluvial samples several metres apart will record the same denudation rate? Secondly, is alluvium collected downstream of
channel junctions in steep terrain mixed sufficiently to represent the entire area upstream? Lastly, what are the affects of landsliding within a basin on the cosmogenic radionuclide concentration measured?

To investigate these issues the importance of sediment mixing in the basin-wide approach to denudation rate measurement will be presented, followed by a brief description of the study setting. The above questions will be considered with appropriate sampling methodologies made explicit. The results obtained will be discussed in terms of assumptions of sediment mixing and the implications in relation to the validity of the technique in rapidly denuding, high relief, terrain will be considered. Finally, the issue of basin size in relation to cosmogenic radionuclide analysis will be briefly discussed.

4.2 Importance of mixing:

Using cosmogenic radionuclides in alluvial sediment in order to derive denudation rates requires several assumptions be valid. Model based assumptions concerning rates of $^{10}$Be and $^{26}$Al production and decay have been dealt with in Chapter 3 (section 3.3). If these assumptions are met, and the basin is in a state of secular equilibrium with respect to the cosmogenic radionuclide concentration, the cosmogenic radionuclide signature in alluvium will correspond to a rate of denudation (Fig. 3.1). However, this will only be true if the alluvial sediments are thoroughly mixed (Bierman and Steig 1996). If sediments are not thoroughly mixed, a sample cosmogenic radionuclide concentration will be biased towards a particular source location within the basin. This location may not have the same rate
of denudation, or indeed cosmogenic radionuclide production, as the whole basin and errors will occur. There are several ways insufficient mixing may occur in small mountainous catchments. Sediment supply to channel reaches from hillslopes may be localised and not representative of the upstream area (Sutherland et al., 2002). Sediments may travel in segregated packets downstream and mix insufficiently at tributary junctions (Miller and Benda, 2000; Benda et al., 2004). Finally, sediment may be unearthed from depth by anthropogenic or landslide activity, diluting the cosmogenic radionuclide concentration of samples by introducing grains not exposed to cosmic rays (Bierman and Steig, 1996). It should be possible to test these issues using the cosmogenic radionuclide concentrations obtained from stream channels with an appropriate sampling strategy.

It is important to remember that, as long as decay is negligible (section 3.3), the problematic issue is one of insufficient mixing and not heterogeneous denudation. So long as sediments from different regions are mixed in proportion to the rate of mass loss from those regions the resulting cosmogenic radionuclide concentration will accurately represent the average basin-wide denudation (section 3.3.2; equation [3.4] and Fig. 3.2).

4.3 Test-site:

The focus of this work was to test assumptions of basin-wide cosmogenic radionuclide analysis in a rapidly denuding, high relief environment. The southern San Bernardino Mountains provide the ideal location, displaying some of the greatest relief and highest rates of long-term denudation in southern California.
(Blythe et al., 2002; for a more detailed description of the mountains see Chapter 5). They are located in the Transverse Ranges which sit astride the ‘Big-Bend’ of the San Andreas Fault Zone. The southern limits of the mountains encroach on the main trace of the San Andreas Fault Zone and consist of two significant, discrete, fault bounded tectonic blocks (Fig. 4.1).

Figure 4.1 A 10 metre resolution digital elevation model showing the southern San Bernardino Mountains comprised of the San Gorgonio and Yucaipa Ridge blocks separated by Mill Creek. The red box in the inset shows the location of the study in relation to the rest of the San Bernardino Mountains.

The Yucaipa Ridge block is a narrow, east-west trending, ridge which has splays of the San Andreas Fault running along the toes of the ridge slopes (Allen, 1957; Spotila et al., 1998). This block has steep slopes and displays much evidence of
rapid denudation in the form of debris chutes and shallow soil slippage (Fig. 4.2) and landslides and debris flow activity have been documented (Davis, 1989; Morton and Hauser, 2001; Tan, 1990).

The San Gorgonio block is an east-west trending antiform displaying >2 km of relief with steep northern and southern slopes. It lies just to the north of Yucaipa Ridge block but is wider, approximately 5 km, and sits around 1 km higher. A notch is cut near the western end of the block which may represent a wind gap where a once active channel traversed the ridge (Spotila, 2001, pers. comm.). The basin formed in the gap is ~20 km², displays ~2 km of relief and is underlain homogeneously by quartz monzonite (Bortugno and Spittler, 1986). There is evidence of localised deep-seated landsliding in this basin (Fig. 4.3a) and a range of slopes and denudational processes (Fig. 4.3b and c, Sadler and Morton, 1989).

4.4 Methods:

In order to test assumptions of sediment mixing, the basin located in the notch, or wind gap, of the San Gorgonio block (as discussed in section 4.3) was divided into several constituent sub-basins which were sampled (Fig. 4.4). To specifically consider the effect of landsliding, five small first order drainage basins on Yucaipa Ridge were also sampled. For the purpose of this study only ¹⁰Be was used but the results apply to all cosmogenic radionuclides used in the basin-wide approach. The sampling methodologies are specific to the assumptions being tested and will be detailed in section 4.5 below.
Figure 4.2 Evidence for rapid mass movement in the southern San Bernardino Mountains.
a. Looking up a debris chute on Yucaipa Ridge block. Width of the chute is around 10 m.
b. Shallow landslide scar on Yucaipa Ridge block, about eight metres in diameter.
Figure 4.3a  Landslide deposits on San Gorgonio block. Large clast in the bottom left is about two metres in diameter. b. Steep lower slopes on the San Gorgonio Block. c. Low relief, upper slopes of San Gorgonio block.
See figure 4.4b

See figure 4.4c
Figure 4.4a  Showing the locations of the basins discussed in the text.  b. Enlargement of the basin sampled on the San Gorgonio block showing sample locations on the drainage network.  c. Enlargement of the basins sampled on Yucaipa Ridge.

All samples consisted of channel sediment, sieved to an appropriate size fraction from which pure quartz separates between ~20 to 60 g were obtained. These were then dissolved with a $^{10}$Be carrier mass ranging from 200-350 µg. The samples were further cleaned of fluorides, B, Fe, Ti, separated from Al and then prepared as targets for accelerator mass spectrometry analysis at the Australian National University (see section 3.6 and Appendix 1). Production rates were scaled using Lal (1991) and muogenic production included according to Stone (2000) and Granger et al., (2001) (section 3.4, Appendix 2 and 3).
4.5 Results and Discussion:

4.5.1 Reach Mixing:

In wide channels draining large basins, or braided rivers, the sediment sampled is usually an amalgamation from across the channel (e.g. Clapp et al., 2001; 2002). Alternatively, it can be assumed that the basin being sampled is large enough that the cosmogenic radionuclide concentration will have had sufficient time during transport to become thoroughly mixed and hence not vary over spatial scales of several metres (e.g. Schaller, 2001). In smaller mountainous basins, sediments may not have sufficient time, or be excavated too infrequently, to allow thorough mixing (Miller and Benda, 2000). This may prove problematic as insufficient sediment mixing, and hence also cosmogenic radionuclide mixing, within the channel over spatial scales of a few metres will cause spurious and inconsistent results. In order to test whether channel sediments in steep San Bernardino Mountain catchments are sufficiently mixed over a scale of several metres, three samples were collected for basin MHC-2 (Fig. 4.4) at five metre intervals. If the sediments were thoroughly mixed the cosmogenic radionuclide concentration of each sample should be similar. Grain sizes less than 2 cm were collected and crushed to 125-250 μm. The constituent quartz grain size is larger than this but could not be used because of potential non-quartz inclusions within these grains (section 3.6.2.1). The results are shown in Figure 4.5 (see data listed in Appendix 3). The nuclide concentrations for the three samples are not within one standard deviation error of each other. However, they are very similar indicating that, while not entirely consistent, sediments are sufficiently mixed to achieve reasonable results. Interestingly, the denudation rates derived from these cosmogenic radionuclide concentrations are
within one standard deviation error of each other (Fig. 4.5). This is due to the error associated with $^{10}$Be production when deriving denudation rates. From this it is surmised that small San Bernardino Mountain basins have the potential to provide sufficient enough sediment mixing to allow the basin-wide approach to be employed with acceptable error. Care must be taken, however, to avoid sites where recent activity may have deposited local colluvial material (Fig. 4.6).

4.5.2 Junction mixing:

4.5.2.1 Mixing model:

When tributaries coalesce, sediments of differing provenance will be mixed. The degree to which these sediments become mixed can result in a sample containing entirely, or predominantly, the cosmogenic radionuclide ‘signature’ of just one of the contributing regions. This will result in a denudation rate estimate based on an erroneous production rate and specific to an unknown location within the basin. In order to test whether this might occur in the San Bernardino Mountains cosmogenic radionuclide samples were collected, in two instances, just downstream of a tributary junction, and from each of the two tributaries upstream of the join. The contributing areas under discussion are shown in Figure 4.7. The cosmogenic radionuclide concentration measured at the downstream sample site should be an amalgamation of the cosmogenic radionuclide concentrations from the two contributing areas, or sub-basins, mixed in proportion to their respective rates of sediment yield. Thus a simple model can be constructed where the cosmogenic radionuclide concentrations leaving the sub-basins can be mixed in proportion to their sediment production. If
Figure 4.5a. Graphs of cosmogenic radionuclide (CRN) concentrations and b. the denudation rates derived from those concentrations all with 1 standard deviation. Note that the radionuclide concentrations are not quite within 1 standard deviation of each other but that the increased error introduced by converting the concentrations to denudation rates means that the three rates agree within error.

Figure 4.6 A debris cone formed by the channel bank near sample site FC. See text for discussion of sampling strategies.
the sediments are thoroughly mixed, the amalgamated cosmogenic radionuclide concentration derived this way will equal the cosmogenic radionuclide concentration recorded in sediment downstream, as shown in equation [4.1], where subscripts 1 and 2 denote separate sub-basins, \( N \) is cosmogenic radionuclide concentration, \( \bar{N} \) is downstream or amalgamated cosmogenic radionuclide concentration, \( \varepsilon \) is denudation rate and \( A \) is the proportion of the whole area that each sub-basin comprises. This methodology is applied to the basins indicated in Figure 4.7 and the two examples will be considered individually below.

\[
\frac{N_1 \varepsilon_1 A_1 + N_2 \varepsilon_2 A_2}{\varepsilon_1 A_1 + \varepsilon_2 A_2} = \bar{N}
\]

[4.1]

4.5.2.2 Mixing model application:

The sample MHC-10 is collected just downstream of a tributary junction, at the head of a basin draining an area composed of two sub-basins, of similar size, underlain by the same quartz monzonite lithology. By collecting sediment just upstream of the junction, the cosmogenic radionuclide concentrations of the two component sub-basins are also sampled, as represented by MHC-11 and MHC-12 (Fig. 4.7a). Sediments are collected and processed as described in sections 4.4 and 4.5.1. with the results given in Figure 4.8a. The graph indicates that, irrespective of the degree of mixing assumed between the sediments from the two sub-basins, the cosmogenic radionuclide concentration of the mixture taken upstream of the junction cannot be
Figure 4.7a. The basins on the San Gorgonio block studied by MHC-12 (orange) and MHC-11 (yellow) are mixed to give the orange shading in figure 4.8a. MHC-10 records the cosmogenic radionuclide concentration given by the green shading in figure 4.8a.  

b. MHC-14 (yellow) and MHC-13 (orange) are mixed to give the cosmogenic radionuclide concentration represented by the orange shading in figure 4.8b. MHC-15 records the concentration given by the green shading in figure 4.8b.
differentiated from the concentration downstream. This is due to both the high error associated with the cosmogenic radionuclide concentrations, in which case estimates as to the degree of mixing may be possible with reduced errors; and because the cosmogenic radionuclide concentrations from the sub-basins are so similar that there is no dominant signature in the downstream sample to trace sediment provenance. A similar methodology to the one applied here was adopted by Clapp et al. (2002) in basins of similar size to this study and by Matmon et al. (2003) in slightly larger basins. However, the similarity in cosmogenic radionuclide concentrations from the sub-basins in these studies also makes it difficult to assign provenance to the sediment delivered downstream.

Sample MHC-15 is collected at the head of a basin draining two sub-basins which are geomorphologically distinct. The sub-basin measured by sample MHC-13 is steep, providing significant relief of ~1900 m and an average elevation of 2300 m, while the sub-basin measured by sample MHC-14 also contains steep slopes, but has only ~600 m relief and an average elevation of 1700 m (Fig. 4.8b). Both basins are 9-10 km² and underlain by identical lithology. In this example the cosmogenic radionuclide concentrations from each sub-basin are significantly different, allowing a dominant signature of one or the other to be apparent in the downstream measurement. As highlighted in Figure 4.8b, the volumes of sediment flux from each sub-basin define a mixing ratio (x-axis) of cosmogenic radionuclide concentrations (y-axis). If these sediments were thoroughly mixed the cosmogenic radionuclide concentration downstream would be a mixture of ~65% from MHC-13 (28±4 x10³ at/g) and ~35% from MHC-14 (42±6 x10³ at/g) as indicated by the blue
Figure 4.8a graphs showing the degree of mixing between sediments from different sub-basins with different radionuclide concentrations (orange shading) and the radionuclide concentration recorded downstream of the confluence (green shading). The orange and green shading represent 1 s.d. error. As can be seen in a, the radionuclide concentrations recorded by MHC-11 and MHC-12 are too similar (or the errors too large) to differentiate how much mixing is occurring at MHC-10. b shows that <35% (red line) of the sediment exiting MHC-13 has to mix with sediment from MHC-14 in order to obtain the radionuclide concentration recorded downstream by MHC-15. However, sediment volumes exiting each basin show that long-term mixing should be ~65% (blue line) from MHC-13 indicating mixing of sediments at the confluence is not thorough.
line. However, the cosmogenic radionuclide concentration downstream, measured by sample MHC-15 (52±6 $\times 10^3$ at/g), is closer to the cosmogenic radionuclide concentration in sediment leaving the sub-basin measured by MHC-14 (i.e. 42±6 $\times 10^3$ at/g). In order to achieve this downstream concentration the actual mixing ratio must be dominated by sediment from MHC-14. This implies that >65% of the sediment sampled downstream of the confluence actually comes from basin MHC-14 and <35% from MHC-13.

These results raise several points for further discussion. In terms of the validity of applying the basin-wide approach in mountainous terrain, it has been shown that sediment from sub-basins of distinctly different cosmogenic radionuclide concentrations may not be sufficiently mixed to obtain an unbiased result. In this case the sample MHC-15 must be composed of a mixture comprising mostly sediment from MHC-14. This is not an entirely unexpected result, as sediment has been noted to travel downstream in discrete packets or pulses (Miller and Benda 2000). A likely scenario is that sediment is flushed from the respective sub-basins inconsistently due to different elevations sustaining differing micro-climates in the San Bernardino Mountains (Minnich, 1986; 1989). The cosmogenic radionuclide samples used here were collected during the spring thaw after heavy winter snowfall and the sub-basin of higher average elevation (MHC-13) receives greater amounts of precipitation giving it a different propensity to transport sediment than the sub-basin of lower average elevation (MHC-14). It would be considered prudent, therefore, when applying the basin-wide approach to cosmogenic radionuclide measurement to select samples far from junctions and measure basins which were as
geomorphologically homogeneous as possible. However, this introduces further points of interest, such as how far downstream would a sample have to be taken to ensure sufficient mixing had taken place, or, how far upstream would one have to go to ensure the basin was geomorphologically homogeneous?

Clearly how long mixing will take and hence how far downstream samples must be taken to ensure thorough mixing will be dependent on the environment under consideration. It may be that in dendritic mountainous drainages, reaches will never be long enough to provide sufficient mixing before another tributary junction is reached. However, the results of reach mixing given in section 4.5.1 would suggest this is not the case. As shown by the first example, using samples MHC-10, MHC-11 and MHC-12, insufficient mixing is only problematic if the radionuclide concentrations being mixed at a confluence from different locations in a basin are significantly different. There should be an optimum range of basin size for mountainous environments which are small enough that slope gradients will be relatively uniform and denudational processes throughout similar but large enough to ensure the maximum degree of sediment mixing occurs, such that the cosmogenic radionuclide concentration measured at the head of the basin is not a function of sediment transport within the basin. As the errors associated with cosmogenic radionuclide analysis are reduced by refined estimates of production rate it will become easier to distinguish cosmogenic radionuclide signatures in downstream samples and application of the basin-wide approach will require more strict sampling strategies in order to benefit from this increased precision.
A final point raised by this work is that in small basins, underlain by uniform lithologies where provenance studies based on isotope signatures from stream discharge (e.g. Pierson-Wickmann, 2000, Bickle et al., 2003) would be inconclusive, the above mixing model has the potential to define sediment source. The downstream cosmogenic radionuclide concentration is a function of the degree of sediment mixing occurring within the stream channel. Therefore, using mixing models such as the one above is not only vital in order to validate the use of the basin-wide technique, it is an application in its own right which has the potential to reveal the degree of mixing which occurs in alluvial sediments at a range of spatial scales.

4.5.3 Landsliding:

There is much geomorphic and documented evidence of landsliding throughout the southern and central San Bernardino Mountains (Morton and Hauser, 2001; Sadler and Morton 1989; Davis, 1989; Tan, 1990; Stout, 1982). Whilst bedrock landsliding might invalidate the use of the basin-wide approach if it were deeper than ~2 to 3 m (Reinhardt, 2003, pers. comm.), shallow landsliding and debris flows are considered to have no measurable effect (Kirchner et al., 2001). However, shallow landsliding or debris flows in a small basin may invalidate the technique by providing a pulse of sediment to the stream channel from only one location within a basin. As highlighted in section 4.5.2, such inconsistent mixing of channel sediment may provide spurious results. Furthermore, the potential exists in landsliding catchments for grain size bias to be introduced (Brown et al., 1995; 1998). Even in the steepest terrain of the Transverse Ranges the denudational processes will be a mixture of
both landsliding and shallower sediment production (Lavé and Burbank, 2004). This may result in a grain size bias whereby only larger grains are produced by landsliding but smaller grains are derived from both granular disintegration and landsliding processes and have more potential to have come from shallower depths where cosmogenic radionuclide production is highest. Brown et al. (1995; 1998) document this effect in a Puerto Rico watershed by measuring the cosmogenic radionuclide concentrations in a variety of sediment sizes. They modelled cosmogenic radionuclide concentrations in the debris avalanching slopes and found they agreed well with concentrations measured in sand sized alluvium but that gravel sized clasts had lower concentrations, probably the result of being excavated from the base of the shallow mass wasting events (~2 m depth).

In order to investigate whether there was a grain size bias in the southern San Bernardino Mountains samples two aliquots of MHC-11 were produced. One comprised of a grain size of 125-250 μm (MHC-11n), the other a grain size of 500-10000 μm (MHC-11g). Cosmogenic radionuclide concentrations of 59±6x10^3 atoms/g and 46±11x10^3 atoms/g were recorded, respectively. While these two results are within one standard deviation of each other, the errors they incorporate are large. This is as a result of relatively high denudation, large carrier mass and small quartz mass (see chapter 3.6.2.2). Furthermore, the basin from which the samples are derived is unlikely to be so steep as to experience landslides deep enough to excavate bedrock from a significant depth. Accordingly, a cosmogenic radionuclide grain size bias cannot be discounted in the San Bernardino Mountain basins incorporating significant landsliding. However, with the exception of one
deep slide on the San Gorgonio block, there is little evidence of deep-seated landsliding in the catchments sampled. There is much evidence for shallow mass wasting on Yucaipa Ridge (Morton and Hauser, 2001), but the bedrock here is highly fractured, unravelling into talus aprons along the base of the slopes and unable to support significant, deep-seated, landsliding (Spotila et al., 2001). The deep-seated bedrock landslide which has occurred on San Gorgonio block (Fig. 4.3) comprises only a small proportion of the total basin area and so would be unlikely to produce enough buried, low concentration, sediment to significantly dilute radionuclide concentrations.

A more significant influence of landsliding when applying the basin-wide approach to denudation rate measurement in the San Bernardino Mountains may result from sampling sediment derived from an unknown location (see section 4.5.2.2). In the small steep basins of Yucaipa Ridge, shallow landsliding, debris flows, mud flows and rockfall constitute a significant portion of the surface processes operating (Morton and Hauser, 2001). Well defined stream channels are rare and migratory and often it is avalanche chutes which transport material to the base of the ridge slopes to be removed by trunk streams. Measuring denudation rates from cosmogenic radionuclide concentrations in what stream channels could be found on Yucaipa Ridge is, therefore, problematic as production rate calculations are integrated across the entire basin but the sediment sampled may derive from only one part of that basin. The five steeper basins experiencing shallow landsliding on Yucaipa Ridge (OC, RSC, UC, FC and SGC) have more than a two fold variation in
measured denudation rates between them (Between 2.7 to 1.2 mm/a, see Appendix 3). Matmon et al., (2003) show an increase in the variance of denudation rates with decreasing basin size for an Appalachian drainage basin, however, their explanation of this as the result of slope or lithology does not appear to apply here, as does any other apparent controlling variable. One explanation for this variation is the stochastic nature of mass movement processes such as shallow landsliding on Yucaipa Ridge. However, the averaging times of the cosmogenic radionuclide denudation rates on Yucaipa Ridge are between 800-2000 years and so such fluctuations might be expected to be smoothed out. An alternative to explain this spread of results may be that the shallow landslip, debris flow and rockfall processes operating in the small, steep, first order drainage basins of the Yucaipa Ridge block (Morton and Hauser, 2001) prevent thorough mixing of sediments, such that the cosmogenic radionuclide concentration in samples collected from the head of the basins may not be representative of the spatially integrated basin cosmogenic radionuclide concentration. However, if the mean denudation rate of several samples was taken it should approximate a mean rate for the area. This would result from some samples having over-estimated and some under-estimated cosmogenic radionuclide production rates. If there were no bias in where sediments were being sourced from, a mean rate would result. Slopes of Yucaipa Ridge are rectilinear, hence lowering uniformly, such that sediment sources would not be expected to be bias towards either high or low elevations in the basins and a mean denudation rate for the ridge, weighted by basin area, can be calculated.

Sample TC, also from Yucaipa Ridge, is not included in this analysis as it does appear to be steep enough to promote the same voracity of surface process observed in the other basins (see Chapter 7).
4.6 Implications:

One of the most pressing concerns of the above analysis is the scale of the basin under consideration. If cosmogenic radionuclide samples are collected from larger basins, sediment mixing becomes less of an issue as there is time for sediments to be fully integrated by the transport process (e.g. Schaller et al., 2000). This effect has been shown by Matmon et al., (2003) in the Appalachian Mountains where sampling progressively downstream revealed a reduction in the variance of the denudation rates with increasing basin area. The size at which basins overcome issues of insufficient mixing will relate to rates of transport processes within the environment under consideration. At some scale the drainage networks of basins may become too short to allow thorough mixing. This problem may be solved by taking the mean of several samples, although in slightly longer reaches, or higher order drainages, this becomes less of a problem as shown by the consistency of the aliquots of MHC-2. In larger dendritic drainage networks, such as the one sampled here on San Gorgonio block, the source of sediment can be traced in order to validate basin-wide cosmogenic radionuclide analysis or as a methodological study in its own right. However, one can envisage sampling progressively upstream, measuring ever decreasing basins, to the point where the degree of mixing becomes insufficient to allow application of the basin-wide approach. At some scale it may be appropriate to use an amalgamation of colluvial samples rather than alluvial material to obtain a spatially averaged rate as the drainage network will not thoroughly mix the mass leaving very small basins (Riebe, 2002, pers. comm.).
As shown in this and the previous chapter, application of the basin-wide approach in mountainous, rapidly denuding, landscapes may resolve many of the issues concerning the validity of assumptions of the cosmogenic radionuclide denudation rate model, such as decay of $^{10}$Be and $^{26}$Al. However, it introduces further complications in terms of sediment mixing assumptions. The above discussion highlights basin scale as a key issue and shows that, while care must be taken to select appropriate basins for analysis, the application of the basin-wide approach is valid in the San Bernardino Mountains. The issue of basin scale and the degree to which the potential for error should constrain interpretation of denudation rates will be considered further in Chapter 8, in light of the results presented in Chapters 6 and 7. In order to appreciate the context of these results a detailed description of the San Bernardino Mountains field area and surrounding regions is given in the following chapter.
5. Geomorphological setting of the San Bernardino Mountains

5.1 Introduction:

The location of the San Bernardino Mountains, adjacent to the ‘Big-Bend’ of the San Andreas Fault in southern California, ties the evolution of the mountains closely to that of the fault zone. An appreciation of the geological history of the San Bernardino Mountains must include the role played by the San Andreas Fault and so a brief synopsis of its origins in relation to the Transverse Ranges will be presented first. This initial section is not intended as a review of the San Andreas Fault but aims to introduce some of the main concepts central to its evolution which provide a framework within which to view the structure and geomorphology of the San Bernardino Mountains. The San Bernardino Mountains appear to have experienced a very different tectonic history to the surrounding mountains of the Transverse Ranges. The second section will deal with this by presenting the structure, geology and current opinions concerning the formation of the San Bernardino Mountains.
Finally, present day topography, surface processes and evidence of the influence of climatic and anthropogenic impact will be considered.

5.2 Evolution of the San Andreas Fault:

5.2.1 Origins of the San Andreas Fault:

The origins of the San Andreas Fault lie in the subduction of the Pacific Plate spreading ridge beneath the western margins of the North American Plate. Oceanic crust to the east of the Pacific spreading ridge, termed the Farallon Plate, was subducted at a trench along the western North American Plate faster than it was being produced so that it shrunk and the spreading ridge propagated towards the trench (Fig. 5.1) (Atwater, 1970; Dickinson, 1981). Around 37 Ma ago the spreading ridge impinged upon the trench, lifting up the continental crust to raise coastal, central and southern California from the sea (Wright, 1991). The continued subduction of the spreading ridge caused a change of the relative plate motions from convergent to transform which consequently formed the proto-San Andreas, the boundary between the Pacific and North American Plates (Dickinson, 1981). The initiation of transform motion on the San Andreas Fault was proposed as no earlier than 30 Ma ago by Atwater (1970) but has since been refined by various authors to between ~23 Ma and ~17 Ma ago (Humphreys and Hagar, 1990). Wright (1991) points out that subduction of the Farallon Plate has been proposed as the mechanism for formation of the Mesozoic batholiths of western United States and arc volcanics of the Sierra Nevada and Peninsular Ranges. Subduction is also thought to have provided fore-arc basins running the length of California that were subsequently filled and elevated, one of these being the Los Angeles basin. However, Wright
(1991) also notes that an alternative model for the development of the western coast of North America argues that the accretion of terrain, translated northwards from Central America, provided the basement of the Coastal and Transverse Ranges. While not incompatible, these two models have yet to be satisfactorily combined.

**Figure 5.1** The Cainozoic evolution of the North American-Pacific plate boundary charting the emergence of the San Andreas Fault Zone, adapted from Wright (1991). Arrows show plate motion. H and Y are Hawaiian and Yellowstone hotspots. L-Los Angeles Basin; F-Farallon Plate; JF-Juan de Fuca Plate; MN-Mendicino fracture zone; MR-Murray fracture zone; NA-North American Plate; SAF-San Andreas Fault. On the 0 Ma reconstruction, major faults and active volcanics are indicated.

5.2.2 Flake tectonics and block rotation:

The flow in the mantle, thought to be providing the horizontal tractive force driving crustal motion and perpetuating slip on the San Andreas Fault, has moved eastwards over time (Hadley and Kanamori, 1977; Humphreys and Hagar, 1990; Wright,
Presently, the Pacific Plate, North American Plate boundary in the mantle is below the Mojave Desert (Yeats, 1981) causing a misalignment of the San Andreas Fault in the mantle with the San Andreas Fault in the crust (Hadley and Kanamori, 1977). This suggests that a regional décollement may exist under a large portion of the Transverse Ranges, a ductile region at depth in the crust allowing 'flakes' or crustal blocks to move on horizontal shear planes (Yeats, 1981; Sylvester, 1988). There is evidence that the flake tectonics model may be appropriate for the Transverse Ranges in that the tectonic blocks making up the range appear to act independently (Hall, 1981; Li et al., 1992). Palaeomagnetic data indicate that the Transverse Ranges experienced mid-Miocene rotation, the western regions exhibiting 55° to 90° clockwise rotation (Luyendyk et al., 1980; Hornafius et al., 1986; Nicholson et al., 1994) while the eastern Transverse Ranges have been subjected to clockwise rotations ranging from 41° (Carter, 1987) to 20-25° (Powell, 1993). A proposed causal mechanism for the rotation of the western Transverse Ranges is the subduction of the Monterey microplate beneath the North American plate around 20 Ma ago acting to resist strike-slip movement of the San Andreas Fault transform and forcing block rotation (Nicholson et al., 1994). Causal mechanisms are not as clear for the eastern Transverse Ranges but may be associated with a 15° anticlockwise rotation of Mojave Desert blocks (Carter et al., 1987). Others have argued for a clockwise rotation of Mojave blocks to accommodate local fault movements at the northern boundary of the Mojave Desert and San Bernardino Mountains (Bird and Rosenstock, 1984), or that the Mojave shows little rotation and acts as a stable 'backstop' for compression of portions of the Transverse Ranges (Weldon et al., 1993). However, it should be pointed out that the
location of the San Bernardino Mountains is not thought to have rotated in the last ~20 Ma (Powell, 1993).

5.2.3 The Big-Bend and the Transverse Ranges:

Where the north-westerly flowing mantle beneath the Pacific plate encounters the deep crustal root of the Sierra Nevada it is deflected to the west and this has resulted in the Big-Bend, a left-step in the right lateral motion of the San Andreas Fault (Fig. 5.2) (Wright, 1991). The formation of the Big-Bend appears to have been accommodated by movement on the Garlock Fault, and indeed it may have created it (Humphreys and Hagar, 1990). The Big-Bend of the San Andreas Fault, on which the Transverse Ranges have evolved, does not conform to the straight singular trace more applicable to the fault in the north. In fact the San Andreas Fault is a term often used as if describing a singular fault which runs continuously southwards from Cape Mendicino to the Gulf of California. In reality the `San Andreas Fault' is actually a composite of many right and left-lateral, normal and reverse faults which interact to display an overall right strike-slip motion when viewed at the macroscale. This composite of faults will be termed the San Andreas Fault Zone, whereas the term San Andreas Fault will be reserved for discussion of the various strands of the San Andreas Fault proper as shown in Figures 5.2 and 5.3.

Since movement began, macroscale strike-slip motion on the San Andreas Fault Zone has propagated eastwards in association with the movement of mantle flow. In southern California this eastwards shift in the locus of slip has resulted in a series of sub-parallel faults from the California Borderland, across the Los Angeles Basin to
Figure 5.2 The Transverse Ranges and surrounding features. Locations are labelled in white and major faults discussed in the text in black. SB Strand and CV Strand indicate the San Bernardino and Coachella Valley Strands of the San Andreas Fault. SGP SO indicates the location of the San Gorgonio Pass step over. Box in inset shows the location of the area of the figure.
Figure 5.3 Showing the crustal blocks and faults of the San Bernardino Mountains. In the white boxes BBB, SGB, YRB, WCB and MB refers to Big Bear, San Gorgonio, Yucaipa Ridge, Wilson Creek and Morongo blocks respectively. NFTS is the North Frontal Thrust System; SATS is the Santa Ana Thrust Fault; ACGF is the Arrestre Canyon Graben Fault; TRF is the Tunnel Ridge Fault and MiCF is the Mission Creek Fault.
the San Andreas Fault Zone (Fig. 5.2). A significant portion of these faults are still active and an appraisal of the total amount of slip which has occurred between the Pacific and North American plates must take into account the movement which has occurred on them all. Major faults in the Los Angeles Basin and Transverse Range region appear to have exhibited reverse, normal and strike-slip behaviour in their lifetime (Wright, 1991) indicating periods of extension and compression (Ingersoll and Rumelhart, 1999). It is transpression across the Big-Bend which is often cited as the causal mechanism for the formation of the Transverse Ranges and the reason for their east-west orientation on a continental margin where all other mountain ranges have a northeast-southwest trend (Atwater, 1970; Wright, 1991; Li et al., 1992; Ingersoll and Rumelhart, 1999; Sadler and Reeder, 1983). However, the mechanisms of vertical crustal movement by transpression are poorly understood (Spotila et al., 1998). Movement on the San Andreas Fault Zone is thought to have shifted from the San Gabriel fault across (what were to become) the San Gabriel Mountains to the current position of the San Andreas Fault Zone in the late-Miocene to early-Pliocene (Crowell, 1979; Ingersoll and Rumelhart, 1999). The timing of this fault relocation coincides roughly with the opening of the Gulf of California and a change of plate motions (Wright, 1991; Ingersoll and Rumelhart, 1999; Humphreys and Hagar, 1990). The relationship between the Gulf opening and the workings of the San Andreas Fault Zone, however, are not clear (Matti and Morton, 1993). Furthermore, it is still unclear what the mechanisms initiating orogenesis of the Transverse Ranges might be. One proposition is that it is a result of the change in plate motions caused by accommodating the kink of the Big-Bend, and this results in far-field stresses producing a zone of transpression (Zoback, 1987). Another
proposal suggests that local mantle convection between the Transverse Ranges and the Salton Sea is causing crustal drag and compression at the Big-Bend (Humphreys and Hagar, 1990). A third hypothesis is that the Transverse Ranges formed as crustal blocks that have rotated between major parallel faults and produced sections of compression and extension along their respective bounding faults (Nicholson et al., 1986). Section 5.3.1.2 discusses mechanisms of formation particular to the San Bernardino Mountains but whichever process is responsible for the formation of the Transverse Ranges as a whole it must account for the lack of a crustal root beneath central portions of the mountains. The lack of a crustal root is indicative of compression and crustal buckling, thought to have begun around the late to mid-Pliocene (Hadley and Kanamori, 1977; Wright, 1991; Humphreys and Hagar, 1990).

Any model of the orogenesis of the Transverse Ranges must also cater for the different evolutionary histories experienced by different parts of the range (Blythe et al., 2002; Spotila et al., 2002; Matti and Morton, 1993).

5.3 The San Bernardino Mountains:

5.3.1 Structure:

5.3.1.1 Faulting in the proto-San Bernardino Mountains:

Before the formation of the modern San Andreas Fault there is stratigraphic evidence of relative uplift between 9.5 and 4 Ma in what was to become the western San Bernardino Mountains (Meisling and Weldon, 1989; Weldon et al., 1993; Cox et al., 2003). The displacement was on the south dipping Squaw Peak Thrust Fault, the surficial trace of which runs partly along the Santa Ana Thrust Fault, and this relative uplift was caused by compression due to the restraining geometry of the San
Gabriel Fault, the focus of strike-slip motion at the time. It ended when movement on the San Gabriel shifted to the modern San Andreas Fault (Weldon et al., 1993).

5.3.1.2 Tectonic structure of the San Bernardino Mountains:

The San Bernardino Mountains are composed of five crustal blocks. These are the Big Bear, San Gorgonio, Yucaipa Ridge, Wilson Creek and Morongo blocks (Fig. 5.3); each has a unique cooling history as demonstrated by thermochronometric studies (Blythe et al., 2000; 2002; Spotila et al., 1998; 2001). The location, style and slip rate of the faults bounding and dissecting these blocks is a major contributing factor to the topographical form and geomorphology of the San Bernardino Mountains and so a discussion of their tectonic structure is presented below. The lithology, formation and geomorphology of these blocks will be considered in greater detail in subsequent sections. The main fault trace of the San Andreas Fault running through the San Bernardino Mountains is termed the San Bernardino Strand, although it is also referred to as the Southern Strand of the San Andreas Fault, with the Mill Creek Fault being termed the Northern Strand. Here the more common nomenclature of San Bernardino Strand and Mill Creek fault will be adopted. Figures 5.2 and 5.3 show the position of these faults and the topographic features discussed in this section.

The portion of the San Andreas Fault Zone that runs along the southern limits of the San Bernardino Mountains, through the San Gorgonio Pass, has been described as a complex structural knot where the main through-going trace of the San Andreas Fault is hard to define (Allen, 1957; Morton and Matti, 1993b; Yule and Sieh, 2003).
This region is characterised by episodes of fault strand development, abandonment and transference of slip (Morton and Matti, 1993b). At the eastern end of the structural knot is a 20 km left step, or stepover, at the latitude of Pinto Mountain Fault which has been developing since the mid-Pliocene. Here, slip is transferred from the San Bernardino Strand to the Coachella Valley Strand of the San Andreas Fault (Fig. 5.2). Both localised compression across this left step and far-field plate motion have been proposed as the mechanisms which are driving uplift of the San Bernardino Mountains (Spotila and Sieh, 2000; Spotila et al., 2001; Matti and Morton, 1993; Morton and Matti, 1993b; Yule and Sieh, 2003). The formation of the San Andreas Fault Zone in the San Gorgonio Pass occurred in the late-Miocene or early-Pliocene, coincident with the abandonment of the San Gabriel Fault. Being initiated at around 4-3.5 Ma ago, the Mission Creek Strand of the San Andreas Fault ran along the southern limit of where the San Bernardino Mountains would rise (Matti and Morton, 1993). At around 1.2 Ma ago, when movement on the Mission Creek Fault was hampered by the growing extent of its aforementioned left step, movement jumped south from the Mission Creek Fault to the newly initiated San Jacinto Fault as a way of circumventing the structural knot (Morton and Matti, 1993a). This caused compression and regional surface uplift of the San Gorgonio Pass area of the order of 700 m (Morton and Matti, 1993b). The Mill Creek Fault Strand of the San Andreas Fault developed around 0.5 Ma ago and partially resolved the structural knot by uniting the northern San Andreas Fault with its southern component, the Coachella Valley Strand, hence reducing the amount of slip on the San Jacinto Fault (Matti and Morton, 1993). The Mill Creek Fault displays high angle, strike-slip movement and shows evidence of dipping 60° to the south, under
the Yucaipa Ridge, but then further east the dip reverses and a separate strand dips 75° to the north (Allen, 1957; Spotila et al., 2001). Clearly this fault also has a significant vertical component to it as indicated by the relief of the Yucaipa Ridge block. Around 125 ka ago, slip on the San Andreas Fault transferred south from the Mill Creek Fault to the San Bernardino Strand. This caused reactivation of the old Mission Creek Fault trace and created the structural configuration present today (Matti and Morton, 1993). The San Bernardino Strand along the southern San Bernardino Mountains is a shallow dipping, oblique, reverse fault with a strike-slip component (Yule and Sieh, 2003). It becomes difficult to define to the south-east where slip is transferred to the near-vertical, strike-slip Coachella Valley segment of the San Andreas Fault via thrusting in the San Gorgonio Pass region (Allen, 1957; Yule and Sieh, 2003). The two strands of the San Andreas Fault in the southern San Bernardino Mountains, the Mill Creek Fault and San Bernardino Strand, bound the Yucaipa Ridge block. Situated in the southern San Bernardino Mountains, the Yucaipa Ridge block consists of a thin crustal sliver and is considered to have the greatest rate of crustal uplift in southern California (Blythe et al., 2002). However, it should be noted that the Mission Creek and Mill Creek Faults are both splays of the more southerly Coachella Valley Fault and that they coalesce at depth (Matti and Morton, 1993; Yule and Sieh, 2003).

The Mill Creek Fault bounds the southern edge of the San Gorgonio block, or San Gorgonio massif, a broad north-south-plunging arch which forms an approximately east-west orientated ridge displaying the greatest elevations in southern California (Yule and Sieh, 2003). To the north this block is probably bounded partially by the
Barton Flats Fault but evidence for this fault is buried beneath what has been described as both a massive landslide deposit (Stout, 1982) and a relict alluvial fan (Sadler and Morton, 1989). Results from thermochronometric studies indicates that this block has undergone post-Miocene tilting ~5° to the west and ~10° to the north (Spotila et al., 1998; 2001). Tilting of strata in the Santa Ana structural low and northward displacement of the Santa Ana river also suggests that this block is being forced north (Sadler and Morton, 1989). The Santa Ana structural low is an intermontane valley through which the Santa Ana River flows. Survival of pre-orogenic sandstone deposits within the valley show that it is not a denudational feature, but formed as the result of uplift of the San Gorgonio and Big Bear blocks on either side (Spotila et al., 1998).

The Santa Ana Thrust Fault is a north dipping thrust that has variable dip along its length, but averages 30-35° (Spotila and Sieh, 2000). Rising above the trace of the fault is the southern escarpment of the Big Bear block which separates the southern edge of this block from the northern edge of the Santa Ana intermontane low. The Santa Ana Thrust Fault originated between the late-Miocene and Pleistocene, as indicated by an unconformable sequence of sedimentary units in the Santa Ana Valley (Sadler, 1982b; Strathouse, 1982; Fig. 5.4). The initiation of uplift of the southern side of the Big Bear block is considered to have pre-dated uplift of the northern flank but the precise timing of uplift is not known. Earliest estimates place the development of the Santa Ana Thrust Fault at the same time as the Miocene Squaw Peak Thrust Fault and suggest that these two are related (Meisling and Weldon, 1989).
Deer Creek Fault runs along the northern edge of the San Gorgonio block before curving northwards, cutting through the Santa Ana valley and the southern escarpment of the Big Bear block. The fault dips down to the north but appears to have a right lateral strike-slip component of ~3-6 km, based on displacement inferred from clast size analysis (Sadler, 1993). The southern escarpment, however, shows only ~1 km horizontal offset implying that the Deer Creek Fault pre-dates the eastern Santa Ana Thrust Fault (Sadler, 1993). Deer Creek Fault, along with another fault splay further to the east, has accommodated much of the westward movement
of the San Gorgonio block as it has slid west to form the southern drainage divide of the Santa Ana basin. The Deer Creek and associated faults may represent a nascent shear zone (Sadler, 1993; Spotila and Anderson, 2004).

The Big Bear block is the largest of the blocks which make up the San Bernardino Mountains and is approximately saddle shaped, delineated by escarpments to the north and south and descending gently to the east, where it impinges upon the Eastern California Shear Zone. The western limits of the block comprise the Western San Bernardino Arch, a northwest-plunging antiform structure that narrows to a point at Cajon Pass. Cajon Pass marks the location of the complex structural transition between the San Bernardino and San Gabriel Mountains through which the San Andreas Fault runs (Bird and Rosenstock, 1984; Meisling and Weldon, 1989). The undulating plateau surface of the block is pertinent to the tectonic geomorphology of the San Bernardino Mountains and will be discussed in more detail in later sections. There appears to have been little internal deformation of the Big Bear block (Spotila et al., 1998; Miller, 1980) and faulting is limited to the Arrestre Canyon Graben and Tunnel Ridge Faults in the northern and western extents of the block. These exhibit a mixture of strike-slip, normal and reverse displacement but are poorly defined and show no significant vertical offset. A mid-Pleistocene age has been ascribed to the most recent movement of these faults (Hart et al., 1988).

The North Frontal Thrust Zone was active from ~2.5 Ma ago, creating the basin of the Old Woman Sandstone (Matti and Morton, 1993; Sadler and Reeder, 1983). It is
an array of thrust and reverse faults dipping between 10° and 70° to the south and east which stretch along the entire length of the northern San Bernardino Mountains (Miller, 1980). It is cross-cut by the Helendale Fault to the east before it becomes obscured within the horst and graben-like topography of the eastern San Bernardino Mountains (Spotila and Anderson, 2004). The North Frontal Thrust Zone has been considered to have become inactive when slip transferred to the dextral Eastern California Shear Zone (Sadler, 1982a; Meisling and Weldon, 1989). However, palaeoseismic study has revealed that portions of it have ruptured within Holocene times and that it is probably still active (Spotila and Anderson, 2004). Studies of anticlinal folds developed in alluvial fans, formed on the piedmont fringing the northern escarpment, indicates that shortening across the North Frontal Thrust Zone is being accommodated in a broad swath, several kilometres in width (Pearce et al., 2004). The northern escarpment developed above the fault system around early Pleistocene when the plateau surface of the northern and central San Bernardino Mountains was raised as an intact block (Meisling and Weldon, 1989), termed the Big Bear block (Spotila et al., 1998).

The subsurface geometry of the faults bounding the Big Bear Block (North Frontal Thrust Zone to the north and the Santa Ana Thrust Fault to the south) has been defined as a ‘mushroom’ or ‘flower’ structure, whereby both faults curve towards each other with depth (Sadler, 1982b). Others consider the block to have been raised on a décollement such that the North Frontal Thrust Zone flattens out with depth, undercutting the entire range till it reaches the trace of the San Andreas Fault to the south (Li et al., 1992; Meisling and Weldon, 1989; Spotila and Sieh 2000). A
‘seismic step’, where the brittle-plastic transition zone in the lithosphere becomes shallower north of the San Andreas Fault Zone, may complicate interpretations further (Yule and Sieh, 2003).

The Morongo and Wilson Creek blocks have also been identified as discrete crustal blocks in several studies (Blythe et al., 2002; Spotila et al., 1998; 2001; Yule and Sieh, 2003). The Morongo block lies in the structurally complicated San Gorgonio Pass region and is bounded to the south by the San Bernardino Strand of the San Andreas Fault (Matti and Morton, 1993). The subtle topographic expression of the Morongo Block may be a function of its youth rather than uplift rate. The Wilson Creek block extends to the west of the Yucaipa Ridge block. While it is probable that the evolution of the Wilson Creek block is tied to that of the Yucaipa Ridge block its formation appears to be more complex, experiencing periods of uplift and subsidence (Spotila et al., 1998; Blythe et al., 2002). Because of these complicating factors neither the Wilson Creek or Morongo blocks were sampled using cosmogenic radionuclide analysis which focused instead on the Big Bear, San Gorgonio and Yucaipa Ridge blocks.

5.3.2 Geology:

The San Bernardino Mountains are principally a crystalline mass containing several confined sedimentary deposits within intermontane basins or in basins around the northern and southern flanks of the range. Here the crystalline and sedimentary lithologies will be considered in turn.
5.3.2.1 Basement:

The San Bernardino Mountains are composed principally of Cretaceous crystalline basement of biotite-rich granodiorite and quartz monzonite which continue northwards into the Mojave Desert (Matti and Morton, 1993; Meisling and Welding, 1982). These lithologies represent batholiths emplaced during the Mesozoic arc volcanics associated with subduction of the Farallon Plate (section 5.2.1) and they intrude into Pre-Cambrian gneiss in the southern portions of the San Bernardino Mountains (Dibblee, 1982). Figure 5.5 displays the lithologies found in the San Bernardino Mountains. The gneiss rocks are coarse grained, laminated quartz diorite gneiss composed of quartz, feldspar and biotite in alternating light and dark laminations. On Yucaipa Ridge this lithology is brittle and highly fractured (Dibblee, 1982; Spotila et al., 2001). Yucaipa Ridge and Wilson Creek blocks incorporate composite lithologies indicative of a source area further to the south (Matti and Morton, 1993; Blythe et al., 2002).

5.3.2.2 Sedimentary deposits:

Sedimentary deposits on the flanks of the northern and western San Bernardino Mountains are the Old Woman and Crowder Formations, respectively. These will be considered first before going on to discuss the formations found in the intermontane Santa Ana and Mill Creek Valleys. Deposits proximal to the southern extents of the range include the San Timoteo Badlands deposits. For the locations of these deposits see Figure 5.6.
Figure 5.5 (continued on next page)
Figure 5.5 (continued) The lithology of the central San Bernardino Mountains in 3D from three different aspects. The white arrows indicate north. Pinks and purples are granite, granodiorites and monzonites. Grey colours are gneiss and mylonite, dark orange is the Baldwin gneiss. Light brown is sandstone. Yellows are undifferentiated alluvium and landslide deposits. Green is Pelona schist and blue is limestone. Faults are indicated by black lines and highways by red dashed lines. a shows a panoramic view with the red box in the inset showing the extent of the images. b is looking south with the northern escarpment and plateau of the Big Bear block in the foreground. c is looking north with the Yuciapa Ridge block and San Gorgonio blocks in the foreground.

The Tertiary deposited Old Woman Sandstone is the key to the structure along the northern range front of the San Bernardino Mountains. It is a buff coloured, cross bedded deposit containing clasts of gneissic and plutonic rocks, with abundant basalt cobbles and boulders of a Mojave Desert provenance (Sadler, 1982a). At the top of the formation there is a sudden influx of marble-cobble conglomerates. These marble deposits have a San Bernardino Mountain provenance and record a change in
Figure 5.6 The location of sedimentary deposits in and around the San Bernardino Mountains as discussed in the text.
source which suggests the beginning of orogenesis of the San Bernardino Mountains (Sadler, 1982a; see section 5.3.3). Along creeks draining the northern range front, the south dipping thrust faults of the North Frontal Thrust Zone over-ride the Old Woman Sandstone and carry the crystalline Big Bear block over the sandstone deposits.

Further to the west the Crowder Formation overlies the Miocene Punchbowl Formation and is overlain by the Harold Formation (Woodburne, 1975; Foster, 1982). The Crowder Formation is 940 m thick and comprised of several units, which are composed of arkosic conglomerate sandstones, siltstones, fine-grained sandstones and coarse-grained fanglomerates (Dibblee, 1982). It has a probable Pliocene age and Meisling and Weldon (1982; 1989) and Foster (1982) suggested deposition between 4 and 2 Ma, which may correlate with deposition of the Old Woman Sandstone (Sadler, 1982a).

Deposits of Santa Ana Sandstone are recognisable for 22 km along the intermontane Santa Ana Valley and in places they are overlain unconformably by Quaternary fanglomerates (Dibblee, 1982) (Fig. 5.4). Santa Ana Sandstone is composed of light grey, coarse grained arkosic sandstones. Lower parts of the formation contain basalt clasts and have a Mojave Desert provenance, with a K-Ar date of 6.2 Ma (Woodburne, 1975). An increase in the amount of gneiss and the introduction of marble clasts indicates the initiation of thrusting on the Santa Ana Thrust Fault (Strathouse, 1982). Sadler (1993) has also identified clasts with a San Gabriel
Mountain provenance in the Santa Ana Valley. In order to deposit these clasts, the San Gabriel Mountains must have migrated west on the San Andreas Fault Zone, past (what was to become) the San Bernardino Mountains, between 3-9 Ma ago. There is a warping down to the east and west and a northward tilting of the Santa Ana Sandstone strata along the length of the Santa Ana Valley. This implies the San Gorgonio block is pushing northward into the valley and the southern escarpment of the Big Bear block (Sadler, 1993).

At the eastern end of the Wilson Creek block is a sandstone deposit that ends at the Wilson Creek Fault, separating the Wilson Creek and Yucaipa Ridge blocks. This is the Mill Creek Formation, while at the western end of the Wilson Creek block the Potato Sandstone forms a thin sliver along the southern range front. The Mill Creek Formation has been deformed into a synclinal structure known as the Mill Creek, or Yucaipa Ridge Syncline (Dibblee, 1982; Sadler et al., 1993). It may be up to 1300 m thick, composed of several highly indurated units of mainly arkosic sandstones and interfingered with micaceous shales and some conglomerate facies (Dibblee, 1982). A Miocene age of 10-13 Ma has been suggested for the Mill Creek Formation, based on plant fossils and a correlation with the timing of deposition of the Punchbowl Formation (Woodburne, 1975; Dibblee, 1982).

The San Timoteo Badlands lie south of the Yucaipa Ridge block. The San Timoteo Formation is deposited here and is a coarse grained sandstone and conglomerate overlying the Mount Eden Formation, which is composed chiefly of sandstones and mudstones (Morton and Matti, 1993a). The timing of the transition from the Mount
Eden to San Timoteo Formation at ~4.6 Ma ago correlates to a change in source from the San Jacinto Mountains to San Gabriel Mountains (Albright, 1999). An upper conglomerate unit of the San Timoteo Formation contains lithologies with a Santa Ana Valley provenance (Morton and Matti, 1993a). Based on a rodent tooth, the upper part of the conglomerate has an age of 1.5 Ma (Albright, 1999; Morton and Matti, 1993a; Repenning, 1987).

5.3.3 Timing of formation of the San Bernardino Mountains:
The San Bernardino Mountains are not typical of the Transverse Ranges and are considered to be the product of a younger period of orogenesis than neighbouring mountain ranges (Blythe et al., 2002; Dibblee, 1982; Sadler and Reeder, 1983; Spotila et al., 2002). The above discussions of tectonic structure and lithology help place constraints on the timing of the creation of the San Bernardino Mountains and here this information is compiled in a discussion which includes the relative, surface and bedrock uplift of the range.

Reconstruction of the drainage patterns around the area which would become the San Bernardino Mountains indicates that at the Miocene-Pliocene boundary the site of the San Bernardino Mountains was a surface of low relief with some raised areas of resistant quartzite (May and Repenning, 1982; Sadler and Reeder, 1983; Meisling and Weldon, 1982). May and Repenning (1982) dated fossil assemblages near the transition from Mojave Desert to San Bernardino Mountain provenance in Old Woman Sandstone, based on correlations of ages with fossils from other localities, at 2 to 3.2 Ma ago. Relative uplift along the North Frontal Thrust Zone post-dates
this. Meisling and Weldon (1892) provide an age for the Crowder Formation (of between 2 and 4 Ma), using fission-tracks in detrital zircon from a tuff near the base of the formation, bolstering the suggestion by Sadler (1982b) that the Old Woman Sandstone and Crowder Formations are correlated. Combined, these data suggest relative vertical movement on the North Frontal Thrust Zone had begun by 2 Ma at the latest and place an early to mid-Pliocene constraint on the earliest estimate of fault initiation. If the upper units of the Santa Ana Sandstone were deposited contemporaneously with the Crowder Formation (Dibblee, 1982), and the Crowder Formation is correlated with the Old Woman Sandstone (Sadler, 1982a), then a Pliocene age for development of the entire Big Bear block is implied. However, relative vertical movement across the Santa Ana Thrust Fault may pre-date the most significant motion on the North Frontal Thrust Zone (Meisling and Weldon, 1989) and central parts of the San Bernardino Mountains may have begun to rise as early as 3.3 Ma ago (Cox et al., 2003). For the San Gabriel Mountains to have deposited sediments in the Santa Ana Valley between 3-9 Ma ago requires the San Gorgonio Block to have maintained a lower relative elevation than the San Gabriel Mountains during this time (Sadler, 1993). The suggestion by Cox et al. (2003) that San Bernardino Mountain sourced deposits in the Mojave Desert represent relative uplift of the San Gorgonio block 3.3 Ma ago, may instead simply represent uplift on the Santa Ana Thrust Fault; the only age constraint is that uplift initiated post 6.2 Ma ago (Meisling and Weldon, 1989). In this scenario uplift of the Big Bear block was not uniform but rose first in the south after 6.2 Ma ago then in the north between two and three million years ago. The timing of formation of the San Gorgonio block is less certain than that of the Big Bear block (Cox et al., 2003; Spotila et al., 1998;
2001) but during deposition of San Gabriel Mountain clast lithologies in the Santa Ana Valley the crystalline mass of the block was likely still buried and located further to the south-east than present (Sadler, 1993).

Surface uplift rates of between 18 and 35 mm/a have been suggested for the San Gorgonio Block for the period 18-10 ka, based on snowline elevation differences over this period (Morton and Herd 1980). While these rates appear very high in relation to the range of uplift rates estimated for the San Gabriel Mountains of 0.5 to 5 mm/a (Lifton and Chase 1992), they are similar to rates estimated for the western Transverse Ranges of 13 to 20 mm/a (Morton and Herd 1980). However, the rapid surface uplift rates of the San Bernardino Mountains may be based on incorrectly dated periods for glaciation of the San Gorgonio Mountain (Owen et al., 2003), and how the surface uplift relates to the vertical movement of the crust adjacent to the San Gorgonio block at this time is uncertain.

In the south of the range, the Mill Creek Formation is originally thought to have been deposited in a strike-slip basin more than 100 km to the south-east and has been displaced north by strike-slip movement of the modern San Andreas Fault (Sadler et al., 1993). As such, it does not record any information on the orogenesis of the San Bernardino Mountains. The San Timoteo Formation, however, supports the arrival of San Bernardino Mountain deposits south of the mountains no earlier than 1.5 Ma ago (Albright, 1999), with San Gabriel sourced lithologies depositing material from 4.6 Ma ago up until this time. This is coincident with the timing of
deposits in the Santa Ana Valley as the San Gabriel Mountains slid past the San Bernardino Mountains (Sadler et al., 1993).

Further constraints of orogenesis come from thermochronometric studies. The range of (U-Th)/He cooling ages for the Big Bear and San Gorgonio blocks suggest they experienced similar cooling histories (Fig. 5.7). However, younger (U-Th)/He ages at lower elevations on the San Gorgonio block imply that it has experienced a greater amount of denudation (Blythe et al., 2000; 2002; Spotila et al., 1998). A modelled thermal history for the Yucaipa Ridge block, reconstructed from apatite and titanite (U-Th)/He thermochronometry, indicates a period of denudation of ~5 mm/a for this block occurred in the early Pleistocene suggesting rapid bedrock uplift of this block during this time (Spotila et al., 2001). In summary, it appears some central portions of the mountains may have been raised along the Santa Ana Thrust Fault, but that

Figure 5.7 Age elevation plot adapted from Spotila et al. (1998) showing the similarity between the Big Bear (BB) block (circled) and the San Gorgonio (SG) block ages at comparable elevations, suggesting similar cooling histories. The younger San Gorgonio block ages infer more denudation of this block has occurred.
the formation of the central and northern San Bernardino Mountains occurred by relative uplift of the Big Bear block along the North Frontal Thrust Zone 2-3 Ma ago. The timing of formation of the San Gorgonio block is not well constrained and may predate the initiation of movement along the North Frontal Thrust Zone. Cooling histories from thermochronometric studies constrain a period of rapid denudation which suggests formation of the Yucaipa Ridge block occurred ~2.7 Ma ago (Blythe *et al.*, 2002, Spotila *et al.*, 2001).

5.4 Climate and vegetation of the San Bernardino Mountains:

5.4.1 Present climate and vegetation:

The climate varies considerably over several tens of kilometres from south to north across the San Bernardino Mountains. The orographic elevation of air masses associated with winter south-westerly storm tracks as they reach the southern range front of the Transverse Ranges means the southern regions of the San Bernardino Mountains experience higher rainfall, and subsequently sustain different vegetation, than the northern regions. Mean annual rainfall is on the order of 500 mm/a along the foot of the slopes of the southern range front and this value increases to around 1000 mm/a along the crests of the range front peaks (Minnich, 1986). The majority of precipitation at high elevation falls as snow, although snow is rare below an altitude of ~1250 m (Minnich, 1989). Precipitation in the south and central portions of the San Bernardino Mountains is heaviest during November to April, and snow may remain in the northern part of the Big Bear block plateau until late spring. The northern side of the Transverse Ranges are in the lee of a rainshadow and the
northern escarpment of the San Bernardino Mountains has a mean annual precipitation of ~200 mm/a (Minnich, 1989). The majority of precipitation here is delivered by convective summer thunderstorms (Hunt and Wu, 2004) and although such thunderstorms are also common in the southern regions they are usually very localised events (Minnich, 1989). The result is a precipitation gradient decreasing northwards. Mean annual temperatures range from 13-21°C at lower elevations on the northern escarpment of the Big Bear block to 4-10°C at high elevations on the San Gorgonio block, such that temperatures vary with elevation and are higher near the Mojave Desert (Miles and Goudey, 1997).

Vegetation types in the San Bernardino Mountains are dependent on elevation (Minnich, 1989). A Mediterranean climate is found at low elevations. Coniferous stands are present at low altitudes in the south of the mountains but chaparral shrublands become more prevalent further north (Fig. 5.8). In the northern extents of the mountains, descending into the Mojave Desert, juniper trees and creosote bush become dominant (Miles and Goudey, 1997) (Fig. 5.9). As altitude increases, Mesic environments become more common due to the increasing precipitation and decreasing temperatures. Pine forests dominate at the highest elevations, where small patches of subalpine species are present (Minnich, 1989).

5.4.2 Palaeoclimate:

In order to understand the geomorphological evolution of the San Bernardino Mountains, any changes in the prevailing climate must be accounted for. Reconstructing the palaeoclimate for the San Bernardino Mountains is difficult due
Figure 5.8 The vegetation in the Santa Ana Valley showing coniferous stands and patches of chaparral in the foreground.

Figure 5.9 The vegetation at low elevations on the northern escarpment is mostly Juniper tree and Creosote shrub. This image is taken looking north and the rucksack in the foreground lies in an ephemeral channel presently choked with grus.
to the uncertainty as to how the rainshadow effect, which exists across the mountains, might be influenced by temporal climatic variation. Furthermore, any reconstruction will be made more complex because tectonic events, such as the passage of the San Gabriel Mountains to the south of the San Bernardino Mountains, will doubtless influence patterns of rainfall and temperature. Here, some of the general climatic changes which have occurred in the Transverse Ranges and Mojave Desert regions over the last several million years are examined, along with a discussion as to how climate patterns might have been influenced by the formation of the San Bernardino Mountains.

The south-western United States are considered to have undergone a mid Pliocene wet period, waning in the late Pliocene and early Pleistocene to be replaced by a more arid climate (Fleming et al., 1994; Smith et al., 1993). Bull (1991) proposed that the cooler annual temperatures and wetter winters of the Pleistocene Mojave Desert were replaced by warmer temperatures and a switch to wetter summers at the start of the Holocene, while others (Spaulding, 1990; Hunt and Wu, 2004; Wang et al., 1996) have suggested that Mojave Desert summers have become drier over the Holocene. However, across the south-western United States, periods of pluvial activity are considered to have occurred throughout the late Pleistocene and Holocene (Bull, 1991; McDonald et al., 1996).

The San Gorgonio block was the only site in southern California to have been glaciated during the Pleistocene and Holocene (Ingle, 1958; Sharp et al., 1959; Owen et al., 2003). This may result from the sensitivity of snowlines in the San
Bernardino Mountains to changes in sea surface temperature (Minnich, 1986), or the greater elevation of this block. The glacial activity was limited to small cirque glaciers mid-way along the length of the ridge on the north flank, although readvances have been recorded at 18-20 ka, 16-15 ka, 12-13 ka and 5-9 ka (Owen et al., 2003). Fluvioglacial deposits have been used in the Santa Ana Valley to indicate a change in sediment source from the north to the south side of the valley (Strathouse, 1982).

Formation of the Transverse Ranges has been suggested as a mechanism which contributed to climatic change in south-western USA during the Pliocene (Winograd et al., 1985). Climate, therefore, is and has been closely linked to topography in the San Bernardino Mountains. The orographic control of precipitation and the potential feedback between precipitation and crustal uplift (Willet et al., 1993) means that the pattern of rainfall will have varied spatially and temporally over the lifespan of the San Bernardino Mountains. The San Gabriel Mountains migrated westward, south of the San Bernardino Mountains along the San Andreas Fault, as recently as the Pliocene. Assuming macroscale atmospheric circulation patterns have not varied significantly over this period this would have placed the entire San Bernardino Mountains in a rainshadow, further complicating palaeoclimatic interpretations (Spotila et al., 2002). It is clear that climate, and hence vegetation, will not have been constant over the lifespan of the San Bernardino Mountains. However, the rainshadow which exists presently has probably been a consistent feature since westward translation of the San Gabriel Mountains and growth of Yucaipa Ridge block (~2.7 Ma ago), while the presence of glaciation on the San
Gorgonio block suggests it has been elevated enough since the late Pleistocene to induce significant precipitation in the southern parts of the mountains. Although a rainshadow has probably existed over much of the lifespan of the San Bernardino Mountains it is less clear how global climatic changes such as glacial, interglacial and pluvial periods have influenced rainshadow intensity.

5.5 Geomorphology of the San Bernardino Mountains:

The strong climatic and tectonic gradients throughout the San Bernardino Mountains have produced a broad range of environments. The surface processes in the mountains vary from rapid mass movement to long-term weathering and saprolite formation. The cosmogenic radionuclide analysis of the rates of geomorphological processes operating in the San Bernardino Mountains presented in Chapters 6 and 7 takes advantage of the north-south tectonic and climatic gradients which have been described above. To reflect the macroscale topography of the range along a north-south transect, the mountains are divided into five provinces and the gross geomorphology of each of these provinces is considered in more detail below. The Yucaipa Ridge and San Gorgonio blocks are considered as individual provinces but the topographic variation of the Big Bear block means it has been divided into three. These provinces are the northern escarpment above the North Frontal Thrust Zone, the southern escarpment above the Santa Ana Thrust Fault and the plateau surface raised between these escarpments. The Wilson Creek and Morongo blocks were not included in this study and their geomorphology will not be discussed.
5.5.1 Big Bear block province:

5.5.1.1 Northern escarpment:

The northern range front of the Big Bear block rises from the southern Mojave Desert and delineates the northern-most boundary of the San Bernardino Mountains (Fig. 5.10). It stretches for 80 km, along the length of the North Frontal Thrust Zone as an escarpment declining in elevation towards the eastern ramp of the Big Bear block several kilometres to the east of the Helendale Fault (Fig. 5.11). At the most northerly point of the escarpment is a gap in the range front where Arrestre Canyon drains the north-western portions of the plateau into the Mojave Desert, while further to the west it curves south, merging with the ‘wing’ of the Western San Bernardino Arch (Fig. 5.3). Mid-way along its length the escarpment attains its maximum relief of ~1000 m. The small basins draining the escarpment typically contain first order ephemeral channels but in a few locations streams on the plateau have been captured by escarpment drainage. A previous stream capture event may explain the notch in the escarpment at Arrestre Canyon, or it may simply result from less uplift of the North Frontal Thrust Zone as the trace of the thrust faults are not obvious here. Alluvial fans fringe the base of the northern escarpment forming a piedmont surface. Between 100 to 500 m of Miocene and Pliocene sediments have been deposited on underlying weathered granitic bedrock and they are topped with Quaternary alluvial fan units and fluvial deposits (Eppes et al., 2002; Pearce et al., 2004). Many of these fans appear to be too large to have been deposited by the small escarpment basins which suggests a previously north tilted Big Bear block as the source for these fans and that deposition was reduced when uplift of the escarpment by the North Frontal Thrust Zone created a new drainage divide (Cox et
The presence of scarps and folds on the fan surfaces highlights the locations of the thrust faults of the North Frontal Thrust Zone (Pearce et al., 2004). The youthful appearance of some of these scarps and folds suggest recent fault activity. However, underlying limestone lithologies, which appear in several localities near the Helendale Fault and eastern extents of the predominantly granitic and gneissic range front, could provide relatively young-looking folds and scarps (Eppes et al., 2002). This is as a result of the formation of resistant petrocalcic soil horizons on limestone cored folds and so precludes interpretations of chronology based on the degree of dissection. Approximately mid-way along the length of the escarpment there are two lobate debris flow deposits, the toes of which may be mistaken for fault scarps (Miller, 1987). Directly east of the Helendale Fault is the large Blackhawk landslide deposit which extends from the escarpment into the Mojave Desert for several kilometres (Fig. 5.11).
Figure 5.11 Digital elevation model of the northern and central San Bernardino Mountains showing the location of features discussed in the text.  1. point at which Deep Creek exits the mountains; 2. Arrestre Canyon; 3. location of debris flow lobes mapped by Miller (1987); 4. Helendale Fault; 5. Blackhawk Landslide deposit; 6. Sugar Loaf Mountain; 7. Deer Creek Fault; 8. Bear Creek; 9. Slide Peak.
Figure 5.12 The gentle slopes of the northern escarpment where Arrestre canyon drains into the Mojave Desert. The rounded corestones in the centre indicate more resistant areas exhumed by surface lowering of surrounding weathered granite.

Geomorphic processes operating along the northern escarpment vary. In many of the basins in the northern-most extents of the escarpment processes appear to be similar to those found in the Mojave Desert where there is evidence for deep weathering of the granitic bedrock (Oberlander, 1972). Along the crest of the escarpment and on basin interfluves, granitic tors are common and the more gentle slopes of basins near Arrestre Canyon are mantled by rounded, exhumed corestones which are absent on the piedmont surface (Fig. 5.11 and Fig. 5.12). Basin floors and slopes are typically mantled with thin soils or grus⁵. This evidence points to granular disintegration of underlying weathered bedrock, which in places is exposed

⁵ Grus refers to crumbled granite formed by the weathering of parent bedrock. It is composed of individual angular quartz and feldspar grains of sand and gravel size and lacks clay or fine sediment. The process by which it is produced is termed grussification.
as resistant tors or rounded corestones. Between Arrestré Canyon and the Helendale fault the escarpment sustains greater relief and steeper slopes but apart from the instances cited above, indications of rapid mass movement processes are absent.

5.5.1.2 Plateau province:

The surface of the Big Bear block is an undulating plateau. The mostly granitic lithology is interspersed with exposures of limestone to the east of Big Bear Lake, and a few areas of resistant quartzite provide the only significant relief on the plateau (~400 m in the cases of Sugar Loaf and Onyx Mountain (Fig. 5.11)). The relief of the rest of the plateau is <100 m over short wavelengths of several kilometers (Spotila et al., 2002).

A weathered granitic horizon as much as several tens of metres deep is present across the plateau (Fig. 5.13). Profiles through preserved sections of the horizon reveal how it grades upwards from bedrock to tightly interlocking corestones which become more rounded and interspersed within a saprolite matrix. In the stricter sense, the term saprolite is reserved for bedrock which has undergone isovolumetric weathering, that is to say weathering which is not accompanied by a change in volume. Here, the broader definition of the term saprolite is used to refer to bedrock which has weathered in-situ, retaining the structure of the parent bedrock, but without the requirement for isovolumetric weathering. In places the saprolite is capped by an argillaceous soil which has undergone extreme mineral decay, although more commonly overlying grussified layers have been stripped away to leave piles of resistant corestones forming tors. The formation of the weathered
horizon is considered to pre-date the formation of the San Bernardino Mountains and as such it can be used as a marker horizon representing the pre-orogenic surficial topography of the region (Spotila, 1999). Because of the importance of this weathered surface in interpretations of the evolution of the San Bernardino Mountains the evidence which points to its pre-orogenic nature will be considered in greater detail.

The weathered horizon formed prior to the relative uplift of the San Bernardino Mountains when the plateau surface would have been contiguous with what is now the Mojave Desert (Oberlander, 1970; Dibblee, 1982; Doyle, 1982; Meisling and Weldon, 1989; Spotila et al., 1998; 1999). Evidence pointing towards the pre-orogenic nature of this horizon includes instances where the Miocene-Pliocene Old Woman Sandstone and the mid to late Miocene Crowder Formation onlaps the weathered surface (Meisling and Weldon, 1989). The Santa Ana Sandstone in the Santa Ana Valley also overlies the weathered surface (Sadler, 1993). In the east of the range, remnants of basalt flows which erupted prior to uplift of the San Bernardino Mountains are preserved on top of the horizon both on the floor of the Mojave Desert and on the plateau surface (Oberlander, 1972; Woodburne, 1975). Rounded quartzite clasts, which were littered throughout the region prior to orogenesis, are preserved now only on top of the weathered horizon along ridge crests (Sadler and Reeder, 1983). They have a Mojave Desert or eastern Big Bear block provenance but could not have been deposited by the present drainage system, further supporting a pre-uplift origin for the weathered horizon (Spotila et al., 2002).
Figure 5.13 A roadcut through the weathered granitic horizon on the plateau of the Big Bear block shows intact structure in the form of intrusions (grey bands) which have been weathered along with the surrounding matrix. Height of this exposure is around 12 m and is particularly fine grained showing few large clasts.

Furthermore, the upper plate of the Santa Ana Thrust Fault has been thrust over a surface of quartzite clasts (Sadler, 1982b). Assuming these quartzite clasts in the Santa Ana Valley mantle the surface of the weathered horizon as they do on the
plateau to the north and San Gorgonio block to the south, faulting should post-date the formation of the horizon.

Slopes on the plateau surface are generally low gradient and geomorphic processes reflect this with formation of saprolite. Climate varies across the plateau with almost a five-fold increase in mean annual precipitation from north to south (Minnich, 1989). The vegetation reflects this change going from pine-series dominated slopes in the south to chaparral in the north. The majority of the plateau drains into the Mojave River through Deep Creek flowing west (Fig. 5.11). Prior to 1.5 Ma ago it would have drained almost the entire plateau, however, a drainage capture event when a tributary of the Santa Ana River breached the southern escarpment has resulted in re-routing of flow through Bear Creek in the south (Cox et al., 2003).

5.5.1.3 Southern escarpment province:

The southern area of the Big Bear block is bounded by the Santa Ana Thrust Fault, a series of north dipping thrust faults carrying crystalline basement over the Santa Ana deposits. The southern escarpment rises above the surficial trace of the fault system and divides the southern rim of the Big Bear block from the Santa Ana Valley (Fig. 5.14). The highest point of the escarpment is near the centre, while at its eastern limit the trace of the fault disappears and the escarpment relief decreases till it is no longer significant. The escarpment definition is also less distinct further west and it eventually merges with the southern flank of the Western San Bernardino Arch that makes up the southern range front of the San Bernardino Mountains near Cajon.
Pass. Mid way along the length of the escarpment is a break where Bear Creek flows into the Santa Ana Valley. The steep gorge cut by Bear Creek is the result of stream capture of plateau drainage by channels on the escarpment incising back into the plateau ~1.5 Ma ago (Cox et al., 2003). This would have caused a sudden increase in discharge with Bear Creek being the result. The strike of Deer Creek Fault, a dextral strike-slip fault which bisects the Santa Ana Thrust Fault at a high angle (Fig. 5.3), is such that it crosses the Santa Ana River and Bear Creek channels and may have contributed to the headward incision of Bear Creek. Movement on Deer Creek Fault has offset the escarpment by several hundred metres so that the position of the escarpment east of the fault lies further to the south. This effect also occurs on the unnamed fault that parallels Deer Creek Fault to the east.

Displacement of clast deposits in the Santa Ana Valley implies that 1-2 km of dextral slip has occurred on Deer Creek and related faults suggesting the Deer Creek Fault may have been active prior to formation of the Santa Ana Fault Thrust (Sadler,

Figure 5.14 The southern escarpment of the Big Bear block looking north over the structural low of the Santa Ana Valley. Note the flat nature of the crest of the escarpment indicating the low relief surface of the Big Bear block plateau.
1993). Drainage basins on the southern escarpment are small, mostly containing straight, low-order channels which drain directly into the Santa Ana River trunk stream. To the west, where the escarpment is less well defined, there are a few second and third order reaches which are more sinuous and basins are larger; especially after the Santa Ana River has turned south to exit at the southern range front (Fig. 5.15).

Around the high central portions of the escarpment relief is ~1300 m and there is evidence for significant landsliding. East of Bear Creek, the Seven Pines landslide deposits are found in the Santa Ana Valley and scarred source slopes are observed on the escarpment, although the exact source area is complicated by movement on the Deer Creek Fault (Fig. 5.11). Just to the west of where Deer Creek Fault crosses the Santa Ana Thrust Fault are several smaller landslide scars and there is evidence of a large slide at the eastern extent of the escarpment beneath Sugar Loaf Mountain (Sadler and Morton, 1989; Davis, 1989). Directly to the west of where Bear Creek joins the Santa Ana River, Slide Peak basin is conspicuous as a large unvegetated scar and is the source area for the Slide Peak landslide (Fig. 5.11 and 5.16). There is a cone of debris at the base of the escarpment where the Slide Peak drainage joins the Bear Creek channel (Fig. 5.17) which created a small natural dammed lake in the 1930s and so the landslide would probably have occurred then (Sadler and Morton, 1989). There is more evidence of landsliding further west along the escarpment (Reeder, 1989). Landsliding and fluvial incision are the main agents of denudation on the southern escarpment. The evidence for mass movement processes is more
Figure 5.15  Digital elevation model of the southern San Bernardino Mountains showing the location of features discussed in the text. 1. wind-gap; 2. glacial cirques; 3. Barton Flats; 4. Mill Creek Jump-off; 5. Exit of Mill Creek at the southern rangefront; 6. Exit of the Santa Ana River at the southern rangefront (note the black blocks are not real features but artefacts of the digital elevation model).
common here than on the northern escarpment. Whether or not this is related to increased fault activity or greater precipitation on the southern side of the Big Bear block is uncertain.

The Santa Ana Valley contains several types of deposit originating from the southern escarpment of the Big Bear block and the northern slopes of San Gorgonio block. The valley also contains sediments sourced in the San Gabriel Mountains from when they lay to the south of what was to become the San Bernardino Mountains, around several million years ago. The Santa Ana Valley contains landslide debris, fanglomerates and fluvial deposits, including those of a fluvioglacial origin.

Figure 5.16 Looking north over the Santa Ana Valley at the southern escarpment of the Big Bear block, directly west of where Bear Creek (in the far right of the picture) drains into the Santa Ana Valley. Slide Peak is in the centre of the picture and the unvegetated scar on its eastern flanks (centre right) is the source of the Slide Peak landslide.
Figure 5.17 The foot of Slide Peak where the landslide debris forms a large talus cone near the confluence with Bear Creek and the Santa Ana River.
Also evident is the Santa Ana Sandstone (section 5.3.2.2). The Santa Ana River channel has incised these deposits leaving scarps where the toes of deposits fringing the flank of the escarpment have been removed and in places fluvially smoothed strath terraces imply rapid fluvial incision (Fig. 5.18). A prominent knickpoint is present several hundred metres upstream of the confluence between Bear Creek and the Santa Ana River (Fig. 5.11). This knickpoint separates a steeper Santa Ana channel reach downstream from a lower gradient reach upstream (Fig. 5.19). There is no evidence as to whether the location of the knickpoint is structurally or lithologically controlled, or whether it is migrating upstream; moreover, clearly defined faults or traces are difficult to identify due to thick sedimentary fill in the Santa Ana Valley. Sadler and Morton (1989) have suggested relative crustal uplift as the mechanism by which the Santa Ana River channel is incising.

5.5.2 San Gorgonio block province:

The northern slopes of the San Gorgonio block drain into the Santa Ana Valley whilst the steeper southern slopes drain into Mill Creek. The block is considered to have undergone tilting of ~10° to the north, based on the gradient of the quartzite mantled weathered horizon found at high elevations on the block. This is further supported by a similar tilting of the isochrones suggested by (U-Th)/He cooling ages (Spotila et al., 1998). The presence of the horizon on both the Big Bear and San Gorgonio blocks suggests that, prior to the formation of the Santa Ana Valley, these blocks were conterminous (Sadler, 1983; Spotila et al., 1998). The smooth east-west ridge crest of the San Gorgonio block is interrupted at Angelus Oaks creating a
Figure 5.18 Fluvially smoothed strath terrace in the Santa Ana Valley downstream of the confluence with Bear Creek. Figure for scale. Several metres above the lower terrace is another which has become more degraded.
Figure 5.19 The gradient of the Santa Ana River through the Santa Ana Valley at 500 m intervals taken from a 10 m digital elevation model shows a prominent knickpoint several hundred metres upstream of the confluence between Bear Creek and Santa Ana River. Also evident is that upstream of the knickpoint channel gradients are lower, see text for discussion.

notch in the ridge (Fig. 5.15). This feature is reminiscent of a wind gap where a once active channel would have traversed the block (Spotila, 2001, pers. comm.). The abandoned channel of the wind gap is now elevated ~400 m above the present trunk stream. The drainage basin formed in notch created by the wind gap drains the western slopes of San Gorgonio Peak and is larger than most of the basins on the block. The sediment mixing in this drainage basin was discussed in Chapter 4.

There are several small cirques high on the northern slopes of the San Gorgonio block ridge which indicate previous glaciations (Ingle, 1958; Sharp et al., 1959; Owen et al., 2003). The presence of the weathered granitic horizon at high elevations on the crest and northern slopes of the San Gorgonio block suggests that pre-uplift topography survives on portions of this block. The areas of subdued
relief, high on the northern side of the ridge, contrast sharply with the steeper slopes present at lower elevations and on the south of the ridge (Fig. 5.20). The geomorphic processes high on the San Gorgonio block are similar to those on the plateau surface, where weathering and saprolite formation precludes a gradual surface lowering revealing rounded corestones in a grus matrix. However, the evidence for past glaciations on the San Gorgonio block makes it probable that weathering processes associated with permafrost have operated here as recently as the mid-Holocene (Owen et al., 2003). Shallow landsliding is typical on lower slopes which are mantled by thin soils and the steeper gradient southern slopes have a thinner soil cover and less vegetation than those in the north. There is evidence that the smooth Barton Flats area on the northern San Gorgonio block slope (Fig. 5.15) may be the debris from a large landslide (Stout, 1982); however, it has also been argued it is a remnant of an alluvial fan (Sadler and Morton, 1989; Sung, 1990). Other relict alluvial fan deposits are evident in the wind gap to the west of Barton Flats (Fig. 5.21). Debris chutes are common on the slopes draining the notch formed in the wind gap and there is evidence for a deep-seated landslide (Bortugno and Spittler, 1986) (Fig. 5.22). The basins draining the slopes of the San Gorgonio block are typically low order but the higher elevation of the San Gorgonio block allows for larger catchments and a more dendritic pattern than is found on the Big Bear block escarpments, especially on the northern slopes where slopes are less steep than southern ones. Many of the channels near the base of the southern slope are narrow bedrock gorges with steep knickpoints several metres high suggesting discontinuous changes in base-level (Fig. 5.23).
Figure 5.20a The steep slopes lower of the San Gorgonio Block contrast with the low relief upper slopes shown in b.
The intermontane valley through which Mill Creek flows contains a significant fill of clasts, derived from the steep slopes of the southern San Gorgonio and northern Yucaipa Ridge blocks (Fig. 5.24). During higher flows this fill is transported and deposited in the wash along the southern range front. Clasts are heaped in bars several metres high along the channel length and terraces are formed where the toes of the talus aprons from the surrounding slopes are incised by the meandering channel (Fig. 5.25). The creek has just undergone a period of down cutting and there is further evidence of incision at Mill Creek Jump-off, at the headwall of the Mill Creek catchment (Sadler and Morton, 1989; Spotila et al., 2001) (Fig. 5.15). Mill Creek is postulated to have attained its current form because of a stream capture event. Incision of less resistant sandstone of the Mill Creek Formation on the southern slopes breached Yucaipa Ridge, allowing drainage to exit the southern rangefront further to the east of where it had previously (Sadler and Morton, 1989).
Figure 5.22 Landslide deposits in the wind gap of the San Gorgonio block. Large boulders are evidence this landslide was deep seated.

Figure 5.23 An incised gorge on the steep southern slopes of the San Gorgonio block. Knickpoints and steep sides are indicative discontinuous of rapid down cutting. The knickpoint in this example is around 6 m high.
Figure 5.24 Looking east up the intermontane Mill Creek Valley. The steep slopes of the southern San Gorgonio block are in the right of the picture and the northern slopes of the Yucaipa Ridge block on the left.

Figure 5.25 A terrace formed where Mill Creek has incised the toes of talus aprons at the base of the slopes of Yucaipa Ridge.
5.5.3 Yucaipa Ridge block province:

Yucaipa Ridge is a narrow, steep-sided, east-west oriented crest which is bound on the north by Mill Creek and to the south by the San Bernardino Strand of the San Andreas Fault. At the south-eastern extents, the trace of the San Bernardino Strand disappears in the San Gorgonio Pass stepover (Allen, 1957; Yule and Sieh, 2003). Rapid mass movement processes appear common on the slopes of this block, which have a thin soil covering. On the northern slopes of the Yucaipa Ridge block the transport of colluvial material into Mill Creek occurs by shallow landslides and debris flows. Deep-seated landslides are rare, probably because of the highly fractured nature of the crystalline bedrock (Morton and Hauser, 2001; Spotila et al., 2001). Shallow soil slippage is observed, as are mud and debris flows on northern slopes (Morton and Hauser, 2001). Some of the steeper small basins on Yucaipa Ridge more closely resemble coalescing debris chutes than incising channels, as mass is transported from the ridge and fluvial channels, where present, are straight and short (Fig. 5.26). The block displays the steepest slopes and greatest relief (around one kilometre) at its centre, but there is similar relief and evidence for landsliding along the entire length (Sadler and Morton, 1989). The mean slope for the entire block is ~32°, measured from 10 m United States Geological Survey digital elevation model data. Fringing southern slopes is an apron of alluvial fans which extend southward into the San Gorgonio Pass. Deep incision of Quaternary fan surfaces on the southern Yucaipa Ridge flank have left areas of low relief, or topographic benches; although in places it is difficult to tell whether the surfaces have been raised by tectonic mechanisms or are the result of incision of the thick sedimentary deposits (Allen, 1957). Landslides and rockfall have also been
Figure 5.26 Looking south at the steep slopes of Yucaipa Ridge. The steep headwalls of the small basins often contain several debris chutes and fluvial channels are short and straight. Note the evidence for landsliding and rockfall near the crest. Mill Creek can just be seen at the bottom of the image.

documented on the southern flank of Yucaipa Ridge (Sadler and Morton, 1989). In summary, the geomorphology of the block indicates it is undergoing rapid denudation.

5.5.4 Anthropogenic influence:
The majority of the San Bernardino Mountains are United States National Forest System land and much of San Gorgonio block and some of the northern Yucaipa Ridge block has been designated as wildlife preserves such that direct anthropogenic influence here is limited. There are several disused highways through the mountains. The presently active routes are limited to Highway 38 which extends up Mill Creek, through the wind gap on the San Gorgonio block, over the eastern end of the Santa Ana Valley and through to Big Bear. There are several dirt tracks throughout the mountains which are used infrequently although the presence of off-road vehicles may cause localised denudation. The major population centres in the
mountains are on the plateau (Fig. 5.27). Big Bear and Lake Arrowhead have populations of several thousand and both maintain artificial lakes. The Big Bear Lake area would once have been a marsh but has been artificially dammed at the entrance to Bear Creek. Other smaller towns with more limited populations include Forest Falls, mid way up Mill Creek; Angelus Oaks, in the wind gap on the San Gorgonio block; Seven Pines in the Santa Ana Valley and Crestline west of Lake Arrowhead. Near Big Bear are several ski-runs operating during the winter to spring snow seasons. Other construction is limited to the small areas of private land around the fringes of the mountains, especially on the southern Yucaipa Ridge block and northern escarpment province.

Figure 5.27 The location of population centres in the San Bernardino Mountains. AL-Arrowhead Lake; AO-Angelus Oaks; BB-Big Bear; CL-Crestline; FF-Forest Falls; SP-Seven Pines.
5.6 Previous geomorphological studies:

Previous geomorphological research in the San Bernardino Mountains provides a necessary context for the cosmogenically derived denudation rate measurements and interpretations of geomorphic processes addressed in later chapters of this study. Most studies in the San Bernardino Mountains have been focused on the structure and geology of the mountains with respect to the San Andreas Fault. The rates of surface processes occurring in the San Bernardino Mountains has received less attention. However, recent work utilising fission-track and low temperature thermochronometry techniques have provided rates of long-term denudation. Furthermore, use has been made of the weathered surface which exists in the mountains as a marker horizon to interpret denudation and relative uplift. Here, the studies which have considered the geomorphic evolution of the San Bernardino Mountains will be discussed paying particular attention to those studies which have used and advanced thermochronometric and geomorphic data to derive denudation rates and those studies which have considered surface processes operating in the San Bernardino Mountains.

5.6.1 Rates of denudation:

5.6.1.1 Thermochronometry:

Several thermochronometric studies have been undertaken in the San Bernardino Mountains. Thermochronometric techniques, such as apatite fission-track dating or (U-Th)/He thermochronometry, record the cooling histories of rocks as they are brought to the earth’s surface by denudation (Summerfield and Brown, 1994). Fission track dating utilises the number and lengths of tracks left by the decay of
\(^{238}\)U in apatite grains to provide an ‘age’ and cooling history for a sample. The tracks are annealed at temperatures between around 60°C to 110°C such that the technique records the history of a sample from crustal depths of several kilometres (Blythe et al., 2002; Gleadow and Brown, 2000). The distribution of the track lengths allows the cooling history of a sample to be modelled. (U-Th)/He thermochronometry measures the retention of the He daughter product of U and Th decay at temperatures below 40°C to 85°C (Wolf et al., 1998). As such it can record the length of time it has taken to remove the upper 1-2 km of crust and from this an estimate of denudation can be made (Spotila et al., 2002). Both these techniques are sensitive to the geothermal gradient which is used and to the degree to which the thermal gradient of the crustal might be perturbed by rapid denudation (Spotila et al., 2001; Stuwe et al., 1994).

Spotila et al. (1998) used the results from 14 apatite (U-Th)/He samples, mostly in the central and southern San Bernardino Mountains to infer the long-term denudation history of the San Bernardino Mountains. This study yielded ages between 64 and 21 Ma on the plateau and ages 56 to 14 Ma on the San Gorgonio block in granites with crystalline ages of ~90 Ma (Spotila et al., 1998). On the Yucaipa Ridge and Wilson Creek blocks, however, ages were between 0.7 and 1.6 Ma. This showed the area of most rapid denudation in the mountains over the long-term is located adjacent to the San Andreas Fault. However, the youth of the San Bernardino Mountains precludes the use of (U-Th)/He to derive rates in all but the most rapidly denuding areas because the bedrock now at the surface contains a thermal history that incorporates both pre- and post-orogenic denudation. That is to
say, the (U-Th)/He technique records denudation of the Big Bear and San Gorgonio blocks over a period of time which pre-dates the onset of orogenesis in the San Bernardino Mountains. As such, any denudation rate derived from these samples will reflect both the mountainous and pre-mountainous environment. In rapidly denuding areas, such as the Yucaipa Ridge block, bedrock uplift has brought samples to the surface rapidly enough that the period of time over which the denudation rate derived is averaged, post-dates the onset of orogenesis. As such, denudation rates measured from the Yucaipa Ridge block are a record of the long-term denudation of mountainous terrain. Geological evidence in the Santa Ana Valley points to a broad up warping of the centre of the San Gorgonio block and indicates that it is tilting to the north (Sadler, 1993). Allowing for the central deformation of the block and applying a 10° northward tilt, Spotila et al. (1998) were able to derive a linear age-elevation profile for this block similar to that of samples taken from the southern escarpment of the Big Bear block (Fig. 5.7). The agreement of their age-elevation profiles suggests the two blocks have had similar cooling histories and supports the theory that the remnants of the weathered horizon, found on both the Big Bear and San Gorgonio blocks, are correlative; even though they are now separated by the Santa Ana structural low.

Spotila et al. (2001) collected 11 apatite and 3 titanite (U-Th)/He samples from the Yucaipa Ridge block to investigate the significance of the San Gorgonio Pass stepover in the formation of the San Bernardino Mountains. Of importance here is that this study produced a modelled, long-term, thermal history of the Yucaipa Ridge block showing rapid denudation of ~5 mm/a occurred during 1.8 to 1.25 Ma.
and that since 1.25 Ma there must have been 1-2 km of denudation. This equates to an average rate of 0.8-1.6 mm/a, assuming a thermal gradient of 30°C/km. How the rate of denudation has been distributed over the 1.25 Ma timeframe was not reconcilable but Spotila et al. (2001) considered that the present geomorphology of the Yucaipa Ridge block suggests continuing rates of rapid denudation. Blythe et al. (2002) further constrained thermal history of the San Bernardino Mountains, presenting a synthesis of thermal histories of the San Gabriel and San Bernardino Mountains and incorporating some unpublished data and that of an earlier study (Blythe et al., 2000). A significant conclusion of this study was that the Yucaipa Ridge block has undergone the most rapid denudation (described as exhumation) in southern California.

5.6.1.2 Incision into a marker horizon:

By mapping the distribution and elevations of the weathered granitic horizon remnants, Spotila and Sieh (2000) were able to interpolate a surface which represented the surface topography that would have existed prior to the formation of the Big Bear block and its uplift relative to the Mojave Desert (section 5.5.1.2). From this they derived the amount of relative uplift which has occurred on the Northern Frontal Thrust Zone faults and Santa Ana Thrust Fault along the northern and southern escarpments of the Big Bear block. Spotila et al. (2002) also used the remnants of the weathered horizon on the plateau and San Gorgonio blocks as a datum indicating minimal denudation since uplift of the mountains. Reconstructing this surface in a similar way that Spotila and Sieh (2000) had, they interpolated the
surface topography prior to the relative uplift of the mountains. The assertion that
the weathered horizon remnants represent a surface of minimal post-uplift
denudation is suggested by onlapping relationships in several localities where older
formations overlie the horizon (section 5.5.1.2). It is further constrained by the
fission track thermal history and old (U-Th)/He ages (Blythe et al., 2000, 2002,
Spotila et al., 1998). To consider the long-term denudation rate in the central and
northern San Bernardino Mountains, Spotila et al. (2002) assumed relative uplift of
the mountains began 2.5 Ma ago, based on the dating of sediment deposition in
formations from around the mountains (section 5.3.3). This age, however, may not
be accurate, and although most workers believe the inception of uplift was 2-3 Ma,
there are a lack of constraints on the timing of the southern escarpment and the San
Gorgonio blocks relative uplift such that this chronology may be flawed (Cox et al.,
2003, see section 5.3.3). Furthermore, Spotila et al. (2002) assume that the
weathered horizon once extended at least as far south as the San Gorgonio block,
and that remnants of the weathered horizon found there have since been separated
from the plateau by the Santa Ana Valley (section 5.6.1.1). They use the weathered
horizon found on the San Gorgonio block to extend the analysis as far south as Mill
Creek. By subtracting the form of the present topography on top of the plateau and
San Gorgonio block from the reconstructed topographical surface, and making
several provisions concerning the amount of denudation which has occurred on the
weathered horizon since formation of the mountains, Spotila et al. (2002) were able
to estimate amounts of vertical incision. Averaging this incision over the last 2.5
Ma gives rates of denudation for the north and central San Bernardino Mountains.
Further incorporating the thermochronometric data from the Spotila et al. (2001)
study allowed the production of a map showing long-term denudation rates throughout the San Bernardino Mountains. However, the results of this map are sensitive to several assumption regarding the timing of uplift and the degree of post-uplift denudation, along with several caveats concerning the nature of uplift and crustal tilting in the San Bernardino Mountains. Recognising this Spotila et al. (2002) placed conservative errors of 50% for the rates derived from the incision of the weathered horizon.

Using the map of denudation rates estimated, Spotila et al. (2002) attempted to correlate rates of denudation with the influence of tectonics, climate, lithology and time. The long-averaging time of the denudation rates made it difficult to separate variables in order to consider them individually, and the coincidence of rapid denudation in locations of both high precipitation and tectonic activity prevents rigorous assessment of their relative control. However, a first order approximation of the long-term, or baseline rate, of denudation across the entire San Bernardino Mountains is achieved in this work.

5.6.1.3 Sediment yields:
Measurements of fluvial sediment loads are being undertaken on the Santa Ana River. However, the guaging stations are not suitably placed to allow measurements of short-term sediment yields from the San Bernardino Mountains. Directly to the west of the San Bernardino Mountains, the San Gabriel Mountains have been the subject of several studies of short term denudation rates. The Los Angeles County Department of Public Works has recorded the volumes of material removed from
behind dams, from debris basins and from reservoirs since the 1940s. This has allowed several studies to investigate the short-term denudation. Rates derived from these studies range from ~0.1 to ~4.5 mm/a (Bull, 1991; Lavé and Burbank, 2004; Scott and Williams, 1978).

5.6.2 Processes of denudation:
Knowledge of the denudation processes operating is fundamental not only for the interpretations of denudation rate measurements, but also for the successful application of the basin-wide approach using cosmogenic radionuclide analysis. Here, the previous studies which have considered denudational processes in the San Bernardino Mountains are reviewed.

Because of the presence of small communities in the San Bernardino Mountains there have been several studies focused on hazard evaluation in landsliding regions. Davis (1989), Tan (1990) and Tan and Giffen (1995) produced maps indicating the propensity of slope failure, debris flows and landsliding around the Forest Falls, Seven Pines, Arrowhead Lake and Big Bear communities (Fig. 5.27). Morton and Hauser (2001) provided a more detailed account of the processes of denudation occurring on the northern flank of Yucaipa Ridge and describe how a substantial amount of mass leaving these slopes results from debris flows, mud flows and shallow landsliding. A comprehensive overview of landsliding in the San Bernardino Mountains was undertaken by Sadler and Morton (1989) who paid particular attention to the Santa Ana Valley deposits, contesting the interpretation of Stout (1982) that the Barton flats region represents a landslide deposit. Reeder
(1989) also focused on the southern escarpment describing two deep-seated landslides that occurred in the Pleistocene and Holocene along the western extents of the escarpment. Minnich (1989) discussed the role of landsliding in relation to precipitation and wildfires throughout the mountains and considered whether the triggers of landsliding were climatic or seismic, while Keefer (1994) apportioned 5% of landsliding in the Transverse Ranges to seismic induced landsliding. Miller (1987) defined the locations of two debris flows on the northern escarpment (Fig 5.10) but does not record any bedrock landsliding other than the Blackhawk Slide deposits.

Combined, the above work describes how the frequency and size of landslides, and other mass movement events, are distributed within the San Bernardino Mountains. In the south, on the Yucaipa Ridge block, rapid mass wasting is common and steep slopes are subject to frequent failure. However, evidence that slides here are deep seated is lacking most probably because the bedrock is too fractured to support significant bedrock slides. One deep-seated but relatively small landslide has been identified at high elevations on the San Gorgonio block (Bortugno and Spittler, 1986). However, it is the southern slopes and lower northern slopes of the San Gorgonio block that appear most prone to processes of rapid mass movement. The southern escarpment of the Big Bear block shows significant landsliding has occurred and large deposits can be identified. However, the five deep seated slides identified on the southern escarpment appear to have occurred since the Pleistocene (Reeder 1989, Sadler and Morton, 1989). While landsliding of the Big Bear block southern escarpment may be of a greater magnitude than on Yucaipa Ridge, it would
also appear to be less frequent. With the exception of the massive Blackhawk Slide, which has been dated at around late Pleistocene, the only evidence for mass movement on the northern escarpment are two debris flow lobes (Miller, 1987). Based on the above studies there appears to be an approximate decrease in the frequency and density of landsliding and rapid mass movement in the San Bernardino Mountains with distance from the San Andreas Fault Zone. The magnitude of the events, however, seems to increase.

There has been much less study concerned with processes of fluvial incision and knickpoint migration in the San Bernardino Mountains. There is evidence that the Santa Ana River is adding an inner gorge to its broad structurally defined valley (Sadler and Morton et al., 1989) and evidence from observations presented in section 5.5.1.3 above suggests that a knickpoint may be migrating upstream. The stream capture and drainage development on top of the plateau have been discussed by Sadler and Reeder (1983), who use clast size analysis of quartzite gravels to identify changes in drainage, and Cox et al. (2003), who map Mojave River deposits to the north of the mountains. From these studies it appears Deep Creek used to drain the entire plateau, leaving the northern range front via a deep gorge which has been in place since the formation of the mountains. However, drainage capture by Bear Creek and incising escarpment basins around the edges of the plateau is reducing the size of the catchment draining through Deep Creek.
5.11 Summary:

The San Bernardino Mountains have developed over the last few million years in response to transpression across the San Andreas Fault Zone. They are composed of several discrete, predominantly crystalline, crustal blocks separated by east-west intermontane valleys which contain sedimentary deposits that have been instrumental in constraining the timing of orogenesis. Pronounced north-south variations in relative uplift, climate and vegetation exist in the San Bernardino Mountains, which contributes towards the variations in the type and rates of geomorphic processes observed. To appreciate the influence each of these factors have had on the development of the mountains, and to address the wider context of the role these factors play in mountain building, the next chapter examines what the controls of denudation in the San Bernardino Mountains are, in relation to the tectonic, climatic and lithological contexts presented above.
6. Factors influencing denudation rates in the San Bernardino Mountains

6.1 Introduction:
The principal mechanisms responsible for maintaining rates of denudation have been long debated in the earth sciences (Milliman and Syvitski, 1992; Summerfield and Hulton, 1994). Tectonically driven base-level change, precipitation and other climatic affects, lithology or basin geometry have all been considered as first order controls over a variety of spatial and temporal scales (Hovius, 2000). However, due to a lack of empirical data in situations where the influence of external factors can be constrained, conclusive evidence has proved difficult to obtain. This ongoing debate as to the relative merits of climate, tectonic mechanisms and lithology in controlling rates of denudation has been re-invigorated recently by a flurry of publications arguing both for, and against, the influence of each (Burbank et al., 2003; Dadson et al., 2003; Reiners et al., 2003; Hartshorn et al., 2002; Wobus et al., 2003; Hodges et al., 2004; Gabet et al., 2004). In this chapter the controls of denudation in the San
Bernardino Mountains will be investigated with specific consideration given to the surficial geomorphic processes operating.

The basin-wide approach is employed to derive denudation rates using cosmogenic $^{10}$Be from thirty drainage basins in the San Bernardino Mountains. The San Bernardino Mountains are located in the tectonically active Transverse Ranges, lying astride the San Andreas Fault in southern California (Chapter 5). They have a broad range of well documented surface processes and experience a strong tectonic and climatic gradient. A relationship between slope gradient, taken from 10 m United States Geological Survey digital elevation model data, and denudation, measured using cosmogenic radionuclide analysis, is recorded in the San Bernardino Mountains. This relationship breaks down as landsliding becomes a significant process facilitating the mass flux from slopes. This allows segregation of basins into those steep enough to be experiencing denudation by landsliding and those undergoing slower denudational processes associated with fluvial incision. The controls of denudation in catchments which experience landsliding are considered separately from those which do not. Results show that, irrespective of denudational process, there is no evidence that mean annual precipitation acts as a control of denudation rates in the San Bernardino Mountains over the $10^2$-$10^4$ year averaging time of the cosmogenic radionuclide analysis. Whatever maintains denudation over these timescales must account for rates which vary over three orders of magnitude throughout the mountains. A tectonic mechanism appears to maintain rapid rates, where slopes are steep and landsliding is prevalent. The changing morphology of
slopes over time is proposed as the most appropriate variable controlling denudation rates on a recently uplifted plateau within the San Bernardino Mountains.

6.2 Geological and geomorphological setting:
Having risen since the late Pliocene as several discrete tectonic blocks, the San Bernardino Mountains comprise the major portion of the Transverse Ranges east of the San Andreas Fault (Fig. 6.1a, b) (Sadler and Reeder, 1983; Meisling and Weldon, 1989; Matti and Morton, 1993; Morton and Matti, 1993; Spotila and Sieh, 2000; Spotila et al., 2001). Each block will be considered individually below, with specific reference made to its geomorphological attributes and denudational processes. For a more detailed discussion of these topics refer to Chapter 5.

The Big Bear block comprises a plateau with steep northern and southern escarpments located above thrust faults (Fig. 6.1c). Denudation along the southern escarpment appears to be facilitated partly through mass movement and landsliding (Sadler and Reeder, 1983; Davis, 1989). Evidence for these geomorphic processes increases mid way along the length of the escarpment where it has been breached by a stream capture event producing Bear Creek (Sadler and Morton, 1989). Once, the plateau would have been drained almost entirely by Deep Creek, flowing out into the Mojave Desert from the north-western edge of the Big Bear block, but now much of this catchment discharge flows through the Bear Creek and into the Santa Ana River (Cox et al., 2003). Just downstream of where Bear Creek joins with the Santa Ana River the presence of fluvially smoothed strath terraces provides evidence for recent rapid fluvial incision (Fig. 6.3). There is little evidence of
Figure 6.1a The Transverse Ranges and surrounding features. Locations are labelled in white and major faults discussed in the text in black. SB Strand and CV Strand indicate the San Bernardino and Coachella Valley Strands of the San Andreas Fault. SGP SO indicates the location of the San Gorgonio Pass step over. Box in inset shows the location of the area of the figure. White box on main figure indicates extent of figure 6.10.
Figure 6.1b 3D digital elevation model showing the central portion of the San Bernardino Mountains comprised mostly of the Big Bear block with northern and southern escarpments, the San Gorgonio block, separated from the Big Bear block by the Santa Ana River and the Yucaipa Ridge block, separated from the San Gorgonio block by Mill Creek. Block boundaries are delineated in red. White dashed line is transect in figure 6.1c below.

Figure 6.1c Schematic drawing of the north south transect indicated in b adapted from Spotila et al. (1998). BBB-Big Bear block; SGB-San Gorgonio block; YRB-Yucaipa Ridge block; MB-Morongo block.
landsiding on the portion of the northern escarpment sampled in this study (Fig. 6.4), which appears to be denuding mostly by the same fluvial incision and weathering processes dominant on the plateau surface. However, further east along the northern escarpment there is more significant relief and evidence for landsliding includes the Blackhawk landslide deposits situated near the eastern terminus of the escarpment (Sadler and Morton, 1989). A deeply weathered granitic horizon is present on the plateau which is thought to have formed contiguous with what is now the Mojave Desert, prior to the formation of the San Bernardino Mountains (Oberlander, 1972; Spotila et al., 1998; 2002). In places this horizon is weathered to a depth of several tens of metres (Fig. 6.5). Where intact, the horizon grades up from bedrock and interlocking corestones to saprolite and is capped by thin argillaceous soils (Spotila et al., 2002). The plateau has an undulating low relief.

Figure 6.2 The southern escarpment of the Big Bear block in the central San Bernardino Mountain. Note the flat crest of the escarpment is the southern extent of the plateau.
Figure 6.3 Fluvially smoothed, or fluted, bedrock strath terraces observed on the Santa Ana River just downstream of the confluence with Bear Creek. The unweathered appearance of the lower terrace adjacent to where the figure is stood in the mid ground suggests recent rapid channel incision.
Figure 6.4.a The western end of the northern escarpment. The three basins in the centre of this photograph were sampled in this study (l-r samples GCR, GC, LLC). b Northern escarpment where exhumed rounded corestones and grussification indicate slow rates of surface lowering.
Figure 6.5a A roadcut showing a section of the deeply weathered horizon, in this case soil capped saprolite which still contains evidence of cross-cutting intrusive structure. Figure for scale. b exposed, rounded corestones typical of where the overlying saprolite of the weathered granitic horizon, found throughout the San Bernardino Mountains, has been stripped off.
surface, underlain principally by quartz monzonite (Fig. 6.6), but it encompasses several areas of quartz diorite, granodiorite and more resistant quartzite, which typically maintain higher elevation. Survival of the weathered granitic horizon on top of the plateau point to a mostly transport limited system where weathering dominates the denudational process.

With over 2000 m of relief, the San Gorgonio block (Fig. 6.1b) displays the highest peaks in the San Bernardino Mountains and the only ones to have experienced glaciation in southern California (Ingle and Moran, 1958; Sharp et al., 1959; Owen et al., 2003). Lying to the south of the Big Bear block, the San Gorgonio block comprises an east-west antiformal ridge (Spotila et al., 1998) with a conspicuously smooth crest (Fig. 6.7). This smooth crest is disrupted near the western limits of the
Figure 6.7 The smooth crest of San Gorgonio Block, here looking south-east at the ridge running east-west. Note the notch in the western end of the block which may represent a wind-gap (Spotila et al 2001, pers. comm.) through which the Santa Ana River would have flowed south.

block by a notch, which may represent a wind-gap through which the Santa Ana River once flowed but which now sits ~400 m higher than the modern channel (Spotila, 2001, pers. comm.). Quartzite deposits with a likely Big Bear block source and remnants of the same weathering horizon found to the north are present at high elevations on the San Gorgonio block and suggest this block was once contiguous with the Big Bear block (Sadler and Reeder, 1983; Spotila et al., 2002). They have since been separated by the structural low through which the Santa Ana River drains and the San Gorgonio block has become more elevated and tilted to the north (Spotila et al., 1998). The southern and steep lower slopes of the San Gorgonio block show evidence of landsliding (Bortugno and Spittler, 1986; Tan, 1990; Sadler and Reeder, 1983; Stout, 1982) and rapid incision, while the higher slopes display remnants of the weathered granitic horizon present on the Big Bear block,
suggesting they have yet to experience the resurfacing which is occurring on lower slopes. Figure 6.8 illustrates these different environments, with lower slopes dominated by shallow mass wasting and low-relief higher slopes showing evidence of grussification and emergence of rounded corestones.

The Yucaipa Ridge block is a steep-sloped crustal sliver situated between two strands of the San Andreas Fault, directly south of the San Gorgonio block. It displays less elevation than the San Gorgonio block and is separated from it by an intermontane valley through which Mill Creek flows (Fig 6.1b). The valley has a fill of clasts, derived from the steep slopes of the San Gorgonio and Yucaipa Ridge blocks. During high flows this fill is transported by Mill Creek and deposited in the wash along the southern range front. Clasts are heaped in bars several metres high along the Mill Creek Valley and terraces are formed where the toes of the talus aprons from the surrounding slopes are removed by the meandering channel (Sadler and Morton, 1989). The thin Yucaipa Ridge block forms a ridge running east-west which is undergoing rapid denudation (Spotila et al., 2001; Blythe et al., 2002) and, unlike the mostly granitic Big Bear and San Gorgonio blocks, it is comprised principally of gneiss bedrock (Bortugno and Spittler, 1986). There is abundant geomorphic evidence of shallow mass wasting, debris flow and rockfall activity (Morton and Hauser, 2001; Tan, 1990; Tan and Giffen, 1995; Fig. 6.9). Landslide activity is common but deep-seated slips are rare due to the highly fractured composition of the bedrock unravelling and being transported via channels and debris chutes to the base of the ridge, where it is periodically removed by Mill Creek (Spotila et al., 2001). The steepest slopes are located half way along the ridge and
Figure 6.8.a The steep slopes lower of the San Gorgonio Block contrast with the low relief upper slopes shown in b.
decrease to the east and west. However, debris flows have been recorded along the eastern end of the northern slopes and pose a significant risk to the small community of Forest Falls (Morton and Hauser, 2001; Tan, 1990; Tan and Giffen, 1995).

A strong rain shadow effect exists across the San Bernardino Mountains as Pacific air-masses are orographically elevated over the steep southern range front. Mean annual precipitation reaches >1000 mm/a at high elevations in the south, falling to <200 mm/a in the northern extents bordering the Mojave Desert (Minnich, 1989). Precipitation is also strongly correlated with elevation and, above ~1500 m, much of the precipitation falls as snow during the rainy season which lasts from around November through April (Minnich, 1986). Convective summer thunderstorms are common along the southern extents of the range but are usually localised (Minnich, 1989). More extensive but less frequent summer storms, rather than winter pressure systems, are recognised as the more likely geomorphic agent in the dry northern extents, where a lack of vegetation will encourage flash-flooding (Minnich, 1989;
Hunt and Wu, 2004; Bull, 1991). Vegetation in the San Bernardino Mountains forms zonal elevation belts such that chaparral at lower elevations is replaced by mixed deciduous forest at moderate elevations. Over 1500-1900 m mixed conifer and subalpine pine. Juniper forests are common in the lee of the rain shadow to the north (Miles and Goudey, 1997).

6.3 Methods:

Alluvial sediment was collected from the mouths of thirty drainage basins spanning a north-south transect across the San Bernardino Mountains (Fig. 6.10). Previous work investigating the controls of denudation over million year timescales has discussed the difficulty of isolating individual variables (Spotila et al., 2002). Accordingly, basins were selected which allowed variables to be controlled, had experienced no glaciation and were geomorphologically and lithologically homogeneous enough to allow sufficient mixing (Chapter 4). Basins which may have been impacted by anthropogenic activity were avoided as much as was possible. Cosmogenic $^{10}$Be targets were prepared from quartz separates of the samples (Bierman et al., 2002, see Appendix 1) and measured by Accelerator Mass Spectrometry at the Australian National University (Fifield, 1999). Production rates were scaled for latitude, elevation and snow cover (section 3.4) and integrated across the basins using 10 m United States Geological Survey digital elevation models, allowing derivation of basin-wide bedrock denudation rates (Brown et al., 1995; Granger et al., 1996; Bierman and Steig, 1996; Appendix 2). The averaging time for these denudation rates varies, depending on the rate recorded, over $10^4$-$10^2$ year timescales (Table 6.1).
Figure 6.10 Showing the location of basins sampled in the central San Bernardino Mountains. Inset, above, illustrates the basins sampled within the 'notch' or wind-gap of the San Gorgonio block (fig. 6.7).
<table>
<thead>
<tr>
<th>Sample ID</th>
<th>Average basin elevation (m)</th>
<th>Average basin latitude (dec deg.)</th>
<th>Basin area (km²)</th>
<th>Average basin slope (°)</th>
<th>Denudation rate (mm/a)</th>
<th>Denudation rate error</th>
<th>Weighted averaging time (ka)</th>
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Table 6.1 Showing the cosmogenic $^{10}\text{Be}$ denudation rate and characteristics of basins sampled as discussed in the text. See figure 6.10 for basin locations. Errors are propagated at 1 standard deviation. Appendix 3 gives further details.

Sample MHC-11mean is the average of MHC-11g and MHC-11n. Sample MHC-2mean is the average of MHC-2a, MHC-2b and MHC-2c. Samples MHC-13i and MHC-8i are the denudation rates derived for the basin areas between sample sites MHC-13 and MHC-8 and between sample sites MHC-8 and MHC-2 respectively.
In the northern and central San Bernardino Mountains there are remnants of a deeply weathered granitic horizon (Oberlander, 1972; Dibblee 1982; Meisling and Weldon, 1982; Sadler and Morton, 1989; Spotila et al., 1998). Miocene basalt deposits (Oberlander, 1972) and Mojave Desert sourced clasts (Sadler and Reeder, 1983) cap the horizon. This, along with the onlapping Miocene Crowder Formation (Meisling and Weldon, 1989), Miocene-Pliocene Old Woman Sandstone (Meisling and Weldon, 1989), and Miocene Santa Ana Sandstone (Sadler 1993), supports a pre-orogenic origin for the weathered surface (Oberlander, 1972; Meisling and Weldon, 1989; Spotila et al., 1998). Spotila et al. (2002) reconstructed the pre-orogenic topography by mapping the elevations of the horizon remnants and, from this data, interpolating a topographic surface. By subtracting the modern topography from this surface they were able to measure the depth of incision which has occurred since orogenesis across the Big Bear and San Gorgonio blocks. Formation of the Big Bear block originated 2-3 Ma, based on on-lapping stratigraphic relationships (May and Repenning, 1982; Meisling and Weldon, 1989) and deposition of San Bernardino Mountains sourced sediments (Albright et al., 1999; Matti and Morton, 1993). Accordingly, Spotila et al. (2002) assumed initiation of orogenesis to be 2.5 Ma and derived a denudation rate map for this period (Fig 6.11). Here, the rates from this map are integrated over the basins on the plateau where the cosmogenic technique has been applied (Fig. 6.10) allowing direct comparison of basin-wide denudation rates averaged over $10^6$ years with the cosmogenic derived rates over $10^4$ years. The denudation rates arrived at by Spotila et al. (2002), from incision into the weathered horizon, were assigned conservative errors of ±50% due to the necessary
interpolative nature of their derivation and assumptions of the relict nature of the weathered horizon.

6.4 Results and Discussion:

6.4.1 Denudational Process:

The mean slope gradient for each basin sampled for cosmogenic $^{10}$Be is derived from a 10 m United States Geological Survey digital elevation model\(^6\). Basins on the northern escarpment and plateau have lower mean slope gradients (<22°), whereas on the southern escarpment, San Gorgonio and Yucaipa Ridge blocks, slope gradients are higher.

\(^6\) See http://gisdata.usgs.gov/NED/default.asp
Plotting denudation against slope gradient for all the basins (Fig. 6.12) reveals a relationship in the crystalline bedrock of the San Bernardino Mountains, which appears to break down above \( \sim 25-30^\circ \).

In order to investigate the relationship between denudation and slope, the denudational processes occurring in each basin are considered and the basins grouped according to their propensity to experience landsliding. For basins on the southern escarpment of the Big Bear block, and the Yucaipa Ridge and San Gorgonio blocks, there are published accounts of landslide events, deposits and hazard assessments (Morton and Hauser, 2001; Sadler and Morton, 1989; Davis,
Field investigation and stereoscopic analysis of United States Department of Agriculture, 1:12000 scale, aerial photography was undertaken and revealed that several basins on the San Gorgonio block are not denuding significantly by landsliding. Accordingly, MHC-2, MHC-11 and MHC-12 are placed within the non-landsliding group. All the basins on the plateau and northern escarpment of the Big Bear block are also included in the non-landsliding group (section 5.6.2). Plotting graphs of slope versus denudation for both the landsliding and non-landsliding groups reveals that a strong linear, or near-linear, relationship exists in those basins where fluvial incision, weathering and grussification dominate surficial processes (Fig 6.13a). In the basins experiencing landsliding, debris flows, rockfall and mass wasting there is no strong relationship between mean basin slope gradient and the rate of denudation (Fig 6.13b). Thus, the break in the slope gradient-denudation rate relationship observed in Figure 6.12 is coincident with a change in denudational process. At one end of the scale are transport-limited basins experiencing saprolite formation and grussification, and at the other are detachment-limited basins experiencing rapid mass wasting processes. The point where the mean basin slope gradient has become steep enough that it no longer relates to denudation corresponds to the emergence of landsliding as a significant denudational process. However, the point at which this transition occurs is badly represented by these results. Subsequently, there is an undefined range of slopes around 25-30° where there are no high denudation rates recorded, which would suggest denudation was occurring rapidly by mass wasting, but rates are no-longer an obvious function of slope. Such undefined basins could belong to either the landsliding or non-landsliding group and as such are assumed to
Figure 6.13.a The slope-denudation relationship in basins with no significant evidence of mass wasting denudational processes.

Figure 6.13.b The slope-denudation relationship in basins where evidence of mass wasting denudational processes such as shallow landsliding, debris flows, mud flows and rockfall are present.
These results, showing a break in the slope-denudation relationship coincident with a change in denudational process, empirically support a model whereby denudation rates are related to slope until they reach angle of repose, cannot steepen further and rates of surface lowering become a function of the rate of local base-level lowering, facilitated through mass wasting processes (Ahnert, 1984; Burbank, et al., 1996; Burbank, 2002; Densmore et al., 1998). Spotila et al. (2001) pointed to the similarity of Yucaipa Ridge slope angles with those of the Himalaya as recorded by Burbank et al. (1996). The angle of repose of slopes in the San Bernardino Mountains of between 30-38° also agrees well with other estimates from the Olympic Mountains (Montgomery and Brandon, 2002) and with measurements of the threshold angle for slopes mantled with a rocky colluvium (Carson and Petley, 1970). Furthermore, the denudation rates of those basins experiencing landsliding agrees with work undertaken in the San Gabriel Mountains to the west, which suggests that rates greater than ~0.3-0.4 mm/a are accommodated by landsliding (Lavé and Burbank, 2004).

Results from the Yucaipa Ridge block further illustrate this point. Five of the six cosmogenic samples (FC, OC, RSC, SGC and UC. Fig. 6.10) taken from the Yucaipa Ridge block have mean basin slope gradients ranging from 31° to 38° and so are likely at their angle of repose. However, one sample (TC. Fig. 6.10), has a lesser mean of 25°, so has not yet reached repose (Table 6.1), and exists in the undefined portion of the graph (Fig. 6.12) such that it could belong to either the landsliding or non-landsliding group. This is reflected in the order of magnitude lower denudation rates measured by sample TC with respect to the other Yucaipa
Ridge block samples. It suggests that the basins at their angle of repose fail by shallow landsliding in response to base level drop which steepens them. Basin TC, not steep enough to be at angle of repose, responds more gradually to such events and likely denudes to some extent by mass wasting events but also by fluvial incision and knickpoint retreat. This proposal is consistent with the observation of stable slopes on the part of Yucaipa Ridge where TC is located (Sadler and Morton, 1989). The slopes at their angle of repose show more sensitivity to base-level drops than slopes which are not. Base-level drop will result from either relative crustal uplift of Yucaipa Ridge or fluvial incision of Mill Creek, the trunk stream running parallel to the base of the northern Yucaipa Ridge slopes, and will push many or all of the basins beyond the mechanical strength of their slopes causing failures. The basins on the Yucaipa Ridge block are underlain by both gneiss and granitic lithologies but there is no apparent relationship between denudation rate and lithology, or slope and lithology. The lack of correlation between the Yucaipa Ridge denudation rates and any controlling variable suggests that when slopes reach their angle of repose, failure occurs stochastically. The trigger for slope failure may result from seismic shaking or from precipitation causing increased loading and pore-water pressure (Hovius et al., 2000; Keefer, 1994; Reeder, 1989). It is the relative merits of climate and tectonics in both landsliding and non-landsliding environments which is turned to now.

It is also possible that a base-level drop experienced by Yucaipa Ridge block would occur by crustal down drop of the surrounding blocks but as denudation is similar on both sides of the ridge and considering the unlikelihood of the blocks on either side dropping synchronously this possibility is not probable. However, any reference to crustal motion in this study implies a relative vertical movement across a fault and does not assume the vectors of motion for this movement in a fixed reference frame.
6.4.2 Influence of climate and tectonics:

6.4.2.1 Non-landsliding basins:

Spotila et al. (2002) recognised a strong correlation between precipitation and denudation rates over $10^6$ year timescales, but pointed out the potential coincidence of this correlation as high precipitation is also associated with active, elevated structures. In order to further investigate the role of precipitation at the $10^2$-$10^4$ year timescale of the cosmogenic radionuclide technique, variables such as climate, lithology and the influence of base-level drop were considered individually. The values for mean annual rain and snowfall, detailed below, were taken from Minnich (1986; 1989). The climatic data used to compare with the cosmogenic-derived denudation is measured over a much shorter period of time ($10^0$-$10^1$ years) than the denudation rates. This is potentially problematic, as climatic regimes have probably varied significantly over the $10^2$-$10^4$ year averaging times of the denudation rates (Bull, 1991). However, the climatic gradient across the San Bernardino Mountains is a function of the interaction of the coastal marine layer and the jet stream, producing orographic precipitation along the southern range front (Minnich, 1986). Assuming the direction of the jet stream has not changed significantly over this period the climatic gradient will have persisted. Therefore, while the intensity of climate experienced by the San Bernardino Mountains has been variable over the last $10^2$-$10^4$ years, the rain shadow effect has probably produced a consistent pattern of higher precipitation in the south, waning northwards.

Similar rates of denudation are observed between an elevated basin (MHC-2mean) ($0.141\pm0.031$ mm/a) experiencing relatively high mean annual precipitation (~850
mm/a) on the San Gorgonio block and a basin of comparable average slope gradient (GC) (0.129±0.015 mm/a) in the much dryer (~250 mm/a) northern Big Bear block (Fig. 6.10, Table 6.1). Basin MHC-2mean contains remnants of the pre-uplift weathered horizon (Spotila et al., 1998), indicating that the significant denudation associated with tectonically induced base-level change that is evident at lower elevations on the San Gorgonio block has not yet migrated as far upstream as this basin. This is also indicated by the geomorphology, as lower slopes are steeper and scarred by landslides (Fig. 6.8a), whereas, higher slopes display evidence for gradual surface lowering in the form of rounded corestones mantling low relief terrain (Fig. 6.8b). As such, the high elevations of the San Gorgonio block are reminiscent of the Big Bear block plateau surface and are assumed to be relatively isolated from any base-level changes which have occurred at the base of the slopes. Most of the precipitation falling at high elevations on the San Gorgonio block falls as snow which, with less immediate infiltration and slow spring melt, is less likely to produce slope failure than extreme rainfall events. This could explain why denudation rates here are not higher (Minnich, 1989). However, comparing MHC-2mean with a basin of similar slope (CP) (0.149±0.015 mm/a) on the Big Bear block plateau (Fig. 6.10, Table 6.1), where mean annual precipitation is high (~750 mm/a) but falls mostly as rain (Minnich, 1989), very similar rates are again observed. This implies neither precipitation, nor the proportion of it falling as rain or snow, exerts a strong influence on denudation rates over ~10^2-10^4 year timescales in the San Bernardino Mountains. Hence, across the strong north-south climatic gradient there is no discernable correlation between mean annual precipitation and rates of denudation. As the basins selected display the greatest possible range of
precipitation in these mountains, it is surmised that mean annual precipitation alone
cannot explain the variation of denudation rates in non-landsiding terrain, at the
spatial scales of the basins sampled and the timescales of cosmogenic radionuclide
analysis.

However, although mean annual precipitation is often considered as a control of
denudation rates (Gabet et al., 2004; Hovius, 2000; Reiners et al., 2003) some have
argued it is the seasonality of precipitation rather than annual volumes which are
most relevant to rates of denudation (Ahnert, 1970; Douglas, 1967; Vandenberghe, 2003). In the arid northern San Bernardino Mountains it is rare summer storms
which do most geomorphic work (Minnich, 1989), whereas, along the southern extents of the mountains winter precipitation falls mostly as snow which has a
dampened infiltration impact due to slow spring melt. Summer thunderstorms are
common in the southern San Bernardino Mountains but are generally localised,
occurring over isolated areas (Minnich 1989). It is difficult to appreciate the effect
of rainfall seasonality due to a lack of relevant information for the San Bernardino
Mountains and because it will have varying impact on differing environments
(Ahnert, 1970; Bull, 1991). In the Mojave Desert it has been proposed that extreme precipitation events do most geomorphic work (McDonald et al., 2003), or that switching from dry to wet climates, typical during El Niño years for example,
increases sediment yields (Viles and Goudie, 2003; Kochel et al., 1997). However,
seasonality also appears to vary over a north-south gradient and, as such, does not
exert significant influence on intermediate-term denudation rates in non-landsiding
terrain across the San Bernardino Mountains.
Vegetation types in the San Bernardino Mountains are strongly coupled to climate and so the influence of vegetation cover appears to be at best a minor variable in controlling denudation in this environment. This finding is supported by work carried out on hillslope stability in the Mojave Desert (McDonald et al., 2003). Furthermore, wildfires increase the amounts of debris stripped from slopes in the Transverse Ranges (Lavé and Burbank, 2004). A strong correlation between vegetation type and the extent of fire-related damage would be expected (Minnich, 1989) but there appears to be no obvious influence of wildfires on denudation between the different ecosystems of the northern and southern San Bernardino Mountains. By increasing the available amounts of flammable material, fire suppression has resulted in more catastrophic fires in this region (Minnich, 1989). Previously, fires would be small and dispersed such that fire-induced denudation would be limited and scattered (Minnich, 1989). The long averaging time of the cosmogenic radionuclide analysis applied here is typically insensitive to anthropogenic increases or decreases of denudation rates, such that the denudation rates should incorporate the influence of naturally occurring wildfires averaged over $10^3$-$10^4$ year timescales.

6.4.2.2 Landsliding basins:

Geomorphic processes such as slope failure are often related to high magnitude, short lived precipitation events, common in the San Bernardino Mountains (Minnich, 1989); hence it is likely landsliding is the result of both climate and seismicity (Hovius et al., 2000; Reeder, 1989), as has been recorded in the San
Gabriel Mountains to the west (Lavé and Burbank, 2004). The denudation rate-precipitation relationships discussed in the above section consider only catchments where landsliding is not the dominant denudational process and so they do not rule out the role of climate in driving landsliding in the steeper catchments. In this section the role of climate and tectonics is considered explicitly in those basins where landsliding is common.

It is useful at this point to make a distinction between those mechanisms which drive denudation rates and those which may trigger a denudational ‘event’ in tectonically active mountains. Studies have recorded the correlation between precipitation and sediment yields in high relief environments (Dadson et al., 2003; 2004). By increasing the shear stress and lowering shear strength, precipitation provokes slope failure Reeder (1989). Keefer (1994) and Lavé and Burbank (2004) note the relative contribution to denudation by precipitation induced landsliding in the Transverse Ranges is more significant than landsliding caused by seismic shaking. However, while precipitation or earthquakes can trigger landslides, for slopes to remain at their threshold angle of stability without retreating requires base-level drop at their toes (Burbank, 2002). Hence, over short timescales precipitation, or seismicity, might cause a denudational event but for such events to persist over the longer-term requires a continuous base-level drop. Prevailing climate, or lithology, may also play a role in controlling the threshold angle at which slopes exist over the long-term (Gabet et al., 2004; Schmidt and Montgomery, 1995). However, once slopes have reached their threshold angle it is base-level drop, via fluvial incision or relative
vertical crustal motion, that are the potential drivers of denudation rates (Densmore et al., 1997).

Assuming the slopes of the Yucaipa Ridge are at their angle of repose, as is suggested by Figure 6.12, and the block is prevented from widening because it is bound by strands of the San Andreas Fault, then any relative base-level drop at the foot of the slopes will result in denudation as the slopes exceed their mechanical strength (Ahnert, 1970; Carson and Petley, 1970). Base-level changes may occur through relative crustal uplift on the bounding faults of the Yucaipa Ridge block in relation to the surrounding topography, or by fluvial incision of the channels at the foot of the ridge (Densmore et al., 1997). Mill Creek drains the valley separating the Yucaipa Ridge and San Gorgonio blocks (Fig. 6.14). However, no trunk stream is present along the southern extents of the ridge (Fig. 6.1). The slopes of both the Yucaipa Ridge block and southern slopes of the San Gorgonio block have average gradients of 32° (based on 10 metre United States Geological Survey digital elevation model data), implying that both are at their angle of repose (Fig. 6.12). The comparatively low denudation rate (0.28±0.03 mm/a) recorded from the southern San Gorgonio block by sample OGC (Fig. 6.10) contrasts sharply with the high rates measured in the steep basins on Yucaipa Ridge. If fluvial incision of Mill Creek was driving denudation rates, by steepening the northern Yucaipa Ridge and southern San Gorgonio block slopes, then denudation rates on both sides of the valley would be similar (Fig 6.15a). Furthermore, while upper reaches of Mill Creek indicate fluvial incision (Spotila et al., 2002), the channel in the valley
contains an appreciable amount of deposited fill material not indicative of a down cutting channel. This evidence suggests that channel incision at the foot of the Yucaipa Ridge block slopes is not driving the high rates of denudation recorded there and so a base level drop, induced by vertical crustal movement, must be maintaining the steepness of Yucaipa Ridge slopes, rather than climate driven fluvial incision (Fig. 6.15b). Therefore, while precipitation events might facilitate landsliding in the southern San Bernardino Mountains, the slopes would fail anyway as they were steepened beyond their mechanical strength by crustal uplift. A tectonically driven base-level drop is proposed to control the rate of denudation in the southern San Bernardino Mountains.
Figure 6.15 An illustration of the lowering of the slopes of the San Gorgonio and Yucaipa Ridge blocks across Mill Creek inferred from cosmogenically derived denudation rates. Arrows denote the amount of incision recorded. All slopes are considered to be at angle of repose. 

a. If denudation on the southern slopes of the San Gorgonio block and the northern slopes of the Yucaipa Ridge block were driven predominantly by fluvial incision of Mill Creek the rates on either side of the Mill Creek valley would be similar. 

b. The rate for the Yucaipa Ridge block northern slopes is the weighted mean of rates from basins OC, RSC and UC. The rate for the San Gorgonio block is from sample SGC. The difference in rates between the slopes illustrated by the different depths incised suggests this is not the case and that vertical crustal motion must be driving the base level drop which is promoting high denudation on the Yucaipa Ridge block.
6.4.3 Influence of lithology:

Basins sampled on the Big Bear and San Gorgonio block were limited to those found in quartz monzonite and on the Yucaipa Ridge block to gneiss and granodiorite. Spotila et al. (2002) find a weak association between denudation rate and lithology in the San Bernardino Mountains but note the potential for coincidence between denudation rates and rock type and that any influence is minor. Other cosmogenic denudation rate studies investigating the role of lithology have found little evidence of it as a significant control at the small basin scale adopted here (Clapp et al., 2000, Riebe et al., 2001), especially in tectonically active orogens (Pinter and Keller, 1991), but show that depth of colluvium exerts a strong control of the weathering rate of underlying bedrock (Small and Anderson, 1999; Clapp et al., 2001). For the landsliding basins, slopes are too steep to retain significant amounts of colluvial material. However, the role played by colluvial mantles in controlling denudation on the plateau surface and northern escarpment of the Big Bear block cannot be discounted. Furthermore, Eppes et al. (2002) discuss how the different sediment composition of soils underlain by different lithologies in the northern San Bernardino Mountains may be controlling surface erodibility, although such contrasts have been avoided here by sampling in only granitic basins.

Schmidt and Montgomery (1995) discussed rock mass strength as a control of relief in active, landsliding environments. However, considering the several fold variation in denudation rates in basins adjacent to each other and underlain by the same lithology (e.g. EFKC and BCT, Fig. 6.10, section 5.3.2), or the similarity of rates in adjacent basins underlain by different lithologies (e.g. UC and OC, Fig 6.10), any
influence of bedrock type, or related properties such as jointing and fracture
frequency, on denudation rates over the timescales of cosmogenic radionuclide
analysis in the San Bernardino Mountains is weak relative to other controls. The
highly fractured nature of Yucaipa Ridge bedrock may, however, be controlling the
frequency of landsliding by facilitating a constant unravelling of the bedrock and
preventing deep-seated landslides.

6.4.4 Slope morphology over time as a control of denudation:
A comparison of denudation rates on the plateau, recorded over different timescales
by the cosmogenic technique and by measuring the depth of incision into a pre-uplift
horizon, reveals a strikingly close correlation and a systematic increase in rates over
more recent timescales. This relationship is investigated further below.

Several basins situated on the plateau of the Big Bear block are measured by
samples CP, CLG, CRD, ARC and GC\(^8\). As discussed in section 6.3 the existence
of remnants of a weathered granitic horizon, formed prior to formation of the San
Bernardino Mountains, allowed Spotila et al. (2002) to map denudation rates in the
San Bernardino Mountains averaged over the last ~2.5 Ma (Fig. 6.11).
Subsequently, these rates were integrated across the basins which had been sampled
for cosmogenic radionuclide analysis. This allowed direct comparison of basin-wide
denudation rates, measured by these two different methods, over timescales of 10\(^6\)
years with those over 10\(^4\) years, respectively (section 7.3). Plotting the denudation

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\(^8\) The basin sampled by GC was collected from a channel draining the northern escarpment but is
included here as most of its area is on the plateau.
rates on the same graph (Fig. 6.16) highlights a strong correlation measured over different timescales. This both bolsters confidence that the rates measured by the techniques are accurate and provokes discussion as to the nature of the relationship. The systematic increase of rates recorded over the $10^4$ year timescale with those recorded over the $10^6$ year timescale is not great (~two-fold). However, of interest is not the amount of difference but the significance of the correlation, which highlights a strong spatial relationship. It is tempting to suggest a climatic cause for such a widespread systematic increase, however, as noted above, climate does not appear to be a significant factor in controlling denudation in the relatively small San Bernardino Mountain basins over $10^3$-$10^4$ year timescales. It is proposed this increase in denudation would be expected without the need to invoke the influence of changing environmental factors such as climate. In the absence of internal deformation, uplift of a plateau on bounding faults and subsequent knickpoint migration inwards from the edge would result in progressively steeper slopes (Hovius et al., 1998). A positive feedback would cause denudation rates to increase with slope gradients over time (Fig. 6.12). In this way it is argued that, subsequent to the initial tectonic input which raised the Big Bear block, it is the morphology of slopes over time that is the primary control of denudation on top the plateau. Barring further tectonic input, rates will increase until a more rugged topography is achieved and the slopes will attain an equilibrium form with respect to local base-level. This model has been proposed for the San Bernardino Mountains by Spotila et al. (2002) who noted the Big Bear block is in a transitional state where drainage is undergoing reorganisation and topography is tending towards a more rugged appearance, similar to that observed in the older San Gabriel Mountains to the west.
Figure 6.16 The relationship between denudation recorded on the plateau over timescales of $10^4$ years using the cosmogenic technique and $10^6$ years by incision into a pre-uplift marker horizon. Rates averaged over the intermediate timescale are approximately twice as high as those averaged over the longer timescale.

These results support this interpretation and suggest it may take several million years for the process of headward knickpoint migration to have significant impact on slopes in regions of broad uplift. The Big Bear block is an example of mountainous topography in an early stage of development which is not often recognised (Hovius et al., 1998). Were the block not so broad it would take less time to achieve such an equilibrium, and so the distance between the block bounding faults could also be seen as the primary control of denudation (Ahnert, 1987; section 8.5.1).

6.5 Summary:

In this study the merits of climate, tectonics and lithology have been reviewed for small basins within the San Bernardino Mountains where mass wasting processes are either present or absent. A relationship between the mean basin slope gradient and the basin-wide denudation rate is revealed which decouples as slopes steepen. The basins not experiencing significant landsliding display a linear, or near-linear,
relationship between the mean slope gradient and denudation rate. Basins with steeper mean slope gradients tend to be those experiencing mass wasting processes, and as slopes increase beyond ~30° they approach their angle of repose and begin to fail stochastically. The variations in climate experienced across the San Bernardino Mountains are not reflected by variations in denudation rates. While an argument for crustal uplift maintaining slopes at their angle of repose and hence maintaining denudation on the San Andreas Fault may not be surprising, it is intriguing that even in parts of the range where the influence of crustal uplift has not yet had significant impact, as evidenced by the preservation of the weathered horizon, there is no climatic control. This may be due to tectonic mechanisms overwhelming others (Phillips, 2003) or because denudation rates of small-scale basins may respond to different controls than a larger catchment would (Matmon et al., 2003). Alternatively, the present climatic gradient may not have persisted over the period of measurement of cosmogenic radionuclide analysis ($10^2$-$10^4$ years), such that the modern day climate is not an appropriate analogue for comparison. However, denudation rates in the San Bernardino Mountains vary over three orders of magnitude, so any interpretations of what mechanisms are driving rates must be able to account for such wide variance. Assuming the climatic gradient has persisted, then neither climate nor lithology appears to exert enough influence to cause the variation found in these mountains. Instead it is tectonically induced base level change which is maintaining the high rates of denudation in the southern extents of the San Bernardino Mountains, whilst the steepening of slopes over time is the first order control of denudation in the absence of tectonic influence. These points raise several further issues concerning the spatial scale at which this study was conducted,
the implications of a lack of climatic control, the significance of comparing
denudation rates averaged over differing temporal scales and how these results relate
to the evolution of the San Bernardino Mountains. These topics will be addressed in
Chapter 8, in light of further study using the cosmogenic technique in the San
Bernardino Mountains which is presented in the next chapter.
7. Denudation over different timescales in the San Bernardino Mountains

7.1 Introduction:

The ability to measure denudation rates over different timescales has the potential to address many outstanding issues in the earth sciences. Tracing the variability of rates of denudation through time not only is necessary for a detailed empirical description of landscape evolution but begs questions of the mechanisms driving those rates. Recent studies have utilised the capability of cosmogenic radionuclide analysis to record denudation rates over intermediate ($10^2$-$10^6$ year) timescales in order to bridge the gap between long-term ($10^6$-$10^8$ year) thermochronometry and short-term ($10^0$-$10^1$ year) sediment yield measurements (Cockburn and Summerfield, 2002). This work has proffered insights into the existence of topographic steady-state (e.g. Pratt-Situala, et al., 2004; Matmon, et al., 2003; Burbank et al., 1996; Vance et al., 2003), the role of anthropogenic influence on denudation rates (e.g. Brown et al., 1995; Schaller et al., 2001; Schaller et al., 2002) and the episodic nature of sediment transport (e.g. Kirchner et al., 2001; Nichols et
al., 2002; Clapp et al., 2000). However, studies have typically relied on apatite and zircon fission-track thermochronometric systems for long-term denudation rate estimates (e.g. Vance et al., 2003; Kirchner et al., 2001; Matmon et al., 2003). The closure temperatures of these systems requires the removal of a depth of crust which has not yet occurred in many young mountains ranges. (U-Th)/He thermochronometry utilises a lower closure temperature than the fission-track technique, such that it records the removal of only the top one to two kilometres of crust and so it can be more applicable to nascent mountain ranges. The presence of long-term denudation rate estimates from low temperature, (U-Th)/He, thermochronometry and incision into a dated marker horizon in the San Bernardino Mountains, southern California (Spotila et al., 1998; 2001; 2002), and application of cosmogenic radionuclide analysis, presents a rare opportunity to compare rates of denudation over intermediate and long-timescales in a mountain range undergoing the early stages of evolution. Furthermore, sediment yield measurements from the nearby San Gabriel Mountains (Lavé and Burbank, 2004) provides short-term denudation rates from an area analogous to the southern San Bernardino Mountains, increasing the range of temporal scales over which to compare denudation. In order to determine rates of denudation on an intermediate timescale in the San Bernardino Mountains, twenty-two basin-wide estimates were derived from cosmogenic $^{10}$Be concentrations in alluvial sediments (Bierman and Steig, 1996). These measurements record rates of denudation over $10^2$ to $10^4$ years (section 3.5). They are compared with long-term rates, averaged over timescales of 1.25 to 2.5 Ma and short-term rates measured over $10^1$ years from the adjacent San Gabriel Mountains.
The results provide a chronology of denudation over the lifespan of the San Bernardino Mountains as they have evolved on the compressive Big-Bend section of the San Andreas Fault (section 5.3.3). Denudation rate results, recorded over both the intermediate and long-term, quantify a decrease with distance from the main trace of the San Andreas Fault implying the focus of crustal shortening in the San Bernardino Mountains occurs along a narrow zone adjacent to the main trace of the fault, and that this has been the case over the lifespan of the mountains. A striking consistency in the measurements of denudation rates over timescales spanning five orders of magnitude is revealed for the parts of the mountains encroaching on the San Andreas Fault Zone. These results prompt points for further discussion, namely: 1. are the different techniques employed here to estimate denudation over a range of timescales recording the same phenomena; and 2. what does the uniformity of rates recorded here reveal about the processes of denudation in the San Bernardino Mountains?

7.2 Geological and tectonic setting:

The San Bernardino Mountains make up a central portion of the Transverse Ranges which run east-west, along the Big-Bend section of the San Andreas Fault in southern California (Fig. 7.1a). They are composed of several discrete, crystalline crustal blocks which have risen independently as a result of transpression across the San Andreas Fault Zone over the last 2-3 Ma, as constrained by facies and provenance changes recorded in several deposits around the perimeter of the range (May and Repenning, 1982; Meisling and Weldon, 1989; Reynolds and Reeder, 1982; Albright, 1999), and thermochronological data (Spotila et al., 1998; 2001).
Figure 7.1a The Transverse Ranges and surrounding features. Locations are labelled in white and major faults discussed in the text in black. SB Strand and CV Strand indicate the San Bernardino and Coachella Valley Strands of the San Andreas Fault. SGP SO indicates the location of the San Gorgonio Pass step over. Box in inset shows the location of the area of the figure. White box on main figure indicates extent of figure 7.4.
Figure 7.1b 3D digital elevation model showing the central portion of the San Bernardino Mountains comprised mostly of the Big Bear block with northern and southern escarpments, the San Gorgonio block, separated from the Big Bear block by the Santa Ana River and the Yucaipa Ridge block, separated from the San Gorgonio block by Mill Creek. Block boundaries are delineated in red. White dashed line is transect in figure 6.1c below.

Figure 7.1c Schematic drawing of the north south transect indicated in b adapted from Spotila et al (1998). BBB-Big Bear block; SGB-San Gorgonio block; YRB-Yucaipa Ridge block; MB-Morongo block.
Denudation rates are measured along a north-south transect through the mountains which incorporates samples from the Big Bear, San Gorgonio and Yucaipa Ridge blocks (Fig. 7.1b). The history, features and structural architecture of these blocks is considered below, for a more detailed discussion see Chapter 5.

The Big Bear block has uplifted along opposing north and south dipping thrust faults, resulting in a raised surface between opposing escarpments (Fig. 7.1b and c). The northern escarpment rises above the Northern Frontal Thrust Zone and delineates the San Bernardino Mountains from the Mojave Desert. At the eastern terminus of this fault system the Big Bear block meets the Eastern California Shear Zone and forms a horst and graben-like topography at lower elevations (Fig. 7.1b) (Spotila and Anderson, 2004). The Western San Bernardino Arch delineates the limits of the San Bernardino Mountains as it follows the trace of the San Andreas Fault and narrows to a point at the complex structural knot of Cajon Pass, separating the San Bernardino and San Gabriel Mountains (Fig. 7.1b) (Bird and Rosenstock, 1984; Meisling and Weldon, 1989). The surface of the central Big Bear block is a low-relief, undulating plateau, underlain principally by monzonite. A conspicuous weathered granitic horizon, as much as several tens of metres deep, is present across the plateau (Fig 7.2). Profiles through preserved sections of the horizon reveal how it grades upwards from bedrock to tightly interlocking corestones, which become more rounded and interspersed within a saprolite matrix. In places it is capped by argillaceous soil exhibiting extreme mineral decay, although more commonly overlying grussified layers have been stripped away to leave resistant corestones and
Figure 7.2 Both photos show a roadcut through a section of the deeply weathered granitic horizon, in this case soil capped saprolite which still contains evidence of cross-cutting intrusive structure shown by the grey bands.
tors (Fig. 7.3). This weathered horizon appears to have formed prior to the formation of the San Bernardino Mountains when the plateau surface would have been contiguous with what is now the Mojave Desert (Oberlander, 1970; Dibblee, 1982; Doyle, 1982; Spotila et al., 1998). Evidence pointing towards the pre-orogenic nature of this horizon includes onlapping Miocene-Pliocene sandstone formations and locations where both it and the sandstone deposits have experienced late Miocene faulting (Meisling and Weldon, 1989). Also, remnants of basalt flows, which were emplaced before the formation of the mountains, are preserved on top of the horizon (Oberlander, 1972; Woodburne, 1975). Furthermore, rounded quartzite clasts, which were littered throughout the region prior to formation of the San Bernardino Mountains, are preserved now only on top of the weathered horizon along ridge crests (Sadler and Reeder, 1983). They have a Mojave Desert or eastern Big Bear block provenance but could not have been deposited by the present drainage system, further supporting a pre-orogenic origin for the weathered horizon (Spotila et al., 2002). The southern edge of the Big Bear block is defined by an escarpment located above the Santa Ana Thrust Fault (Fig. 7.1b). The escarpment has been breached mid way along its length by a drainage capture event causing much of the plateau to drain south through the Santa Ana River where once it would have flowed north into the Mojave through Deep Creek on the north west of the Big Bear block (Cox et al., 2003). The Santa Ana River flows through an intermontane structural low which separates the Big Bear and San Gorgonio blocks (Fig. 7.1b).

The San Gorgonio block is more elevated than the Big Bear block. Remnants of the weathered granitic horizon are present on the northern slopes as are quartzite
clasts which, including the thermal histories derived from (U-Th)/He, suggest that this block was once conterminous with the Big Bear block but has since experienced relative uplift and has been tilted to the north (Spotila et al., 1998; Sadler and Reeder, 1983). The San Gorgonio block is an east-west trending ridge exhibiting the highest peaks in southern California and it displays over two kilometres relief. There is evidence that some of these peaks were glaciated during the late Pleistocene and into the Holocene (Ingle and Moran, 1958; Sharp et al., 1959; Owen et al., 2003). The Barton Flats Fault, bounding the block on its northern side (Fig 7.1c), is hidden by alluvial deposits and what are either the remnants of a large landslide (Stout, 1982) or an alluvial fan (Sadler and Morton, 1989). The Mill Creek Fault bounds the steep southern side of the block.
The Mill Creek Fault separates the San Gorgonio and Yucaipa Ridge blocks and was
the active strand of the San Andreas Fault Zone during the Late Pleistocene (Matti
and Morton, 1993). It is a high angle strike-slip fault which shows evidence of
dipping 60° to the south under Yucaipa Ridge, however, further east the direction of
dip alternates and a separate strand dips 75° to the north beneath the San Gorgonio
block (Allen, 1957; Spotila et al., 2001). Clearly this fault also has a significant
vertical component to it as implied by the presence of Yucaipa Ridge. The Yucaipa
Ridge block (Fig. 7.1b) is a narrow, steep sloped, crustal sliver between the Mill
Creek Fault in the north and the San Bernardino Strand of the San Andreas Fault in
the south (Fig. 7.1c). Thermochronometric data suggests that the Yucaipa Ridge
block is experiencing the highest rate of denudation in southern California (Blythe et
al., 2002). The San Bernardino Strand is the presently the main splay of the San
Andreas Fault in the San Bernardino Mountains (Matti and Morton 1993).
However, as it approaches the eastern end of the Yucaipa Ridge block, there is a
stepover about 25 kilometres to the east where slip is transferred to the Coachella
Valley Strand of the San Andreas Fault (Fig. 7.1a). Both localised compression
across this left step and far-field plate motion have been proposed as the
mechanisms which are driving the relative crustal uplift of the San Bernardino
Mountains (Spotila and Sieh, 2000; Spotila et al., 2001; Matti and Morton, 1993).

7.3 Methods:

7.3.1 Denudation rates over the intermediate-term:

The applicability of using cosmogenic $^{10}$Be concentrations from channel alluvium to
derive basin-averaged denudation rates over timescales of $10^2$ to $10^6$ of years has
been shown for a range of environments (Brown et al., 1995; Granger et al., 1996; Riebe et al., 2000; Clapp et al., 2000; Kirchner et al., 2001; Schaller et al., 2001; Vance et al., 2003). Here, twenty-two alluvial sediment samples were collected from channels draining small (0.3-8.4 km$^2$) basins, across a north-south transect spanning the San Bernardino Mountains (Fig. 7.4). Etched quartz separates were extracted using acid digestion techniques and samples were spiked with a $^9$Be carrier. Samples were cleaned of Ti, Fe and B, separated from Al and analysed by the Accelerator Mass Spectrometry facility at the Australian National University (Fifield 1999, see Appendix 1). Cosmogenic $^{10}$Be concentrations were used to derive rates of denudation for each basin employing the production rate scaling of Lal (1991) and Stone (2000), muogenic production rates and denudation rate model of Granger et al. (2001) and corrections for snow and slope shielding as detailed in section 3.4. (Appendix 2)

The cosmogenic $^{10}$Be concentration produced in surficial bedrock by cosmic radiation can be used to model the rate at which mass is being lost from the surface, averaged over the time it takes to denude a depth of bedrock equivalent to the absorption mean free path of the incoming radiation (Lal, 1991). The nucleogenic, fast and slow muogenic components of cosmic radiation produce cosmogenic $^{10}$Be at different depths in bedrock and subsequently each incorporates a different averaging time. In slowly denuding terrain, a large proportion of the isotopes produced by muons at depth will have decayed by the time they reach the surface, meaning most production (~97%, Stone, 2000) occurs by nucleogenic reactions (spallation) within a few metres of the surface. In rapidly denuding terrain, however, cosmogenic $^{10}$Be
Figure 7.4 Showing the location of basins sampled in the San Bernardino Mountains by cosmogenic radionuclide analysis. Rates are given in table 7.1.
Table 7.1 Showing the rates of denudation over long and intermediate timescales and mean basin slopes. The error of the mean denudation rates are weighted means. Intermediate-term denudation rates derived from cosmogenic $^\text{10}$Be as discussed in the text. $\alpha$ denotes long-term denudation rates derived using (U-Th)/He thermochronometry. $\beta$ denotes long-term rates derived using incision into a dated surface. For further details see Appendix 3.
produced by fast and slow muons will have less time to decay before reaching the surface, and so production by both spallation and fast and slow muogenic interactions becomes significant (Granger et al., 2001). Here, production by fast and slow muons are accounted for using the formulation of Granger et al. (2001) to calculate denudation rates, so the averaging time must reflect the different absorption mean free paths of both muogenic and nucleogenic production pathways. In order to do this the proportion of $^{10}$Be radionuclides produced by spallation, fast and slow muons were calculated and used to weight the mean averaging time of each sample (Table 7.1). It should also be noted, however, that because cosmogenic radionuclide production increases exponentially towards the surface, the denudation rates recorded will be bias towards the present (section 3.5).

7.3.2 Denudation rates over the long-term:

The longer-term pattern of denudation rates in the San Bernardino Mountains are obtained using a combination of (U-Th)/He thermochronometric data from the Yucaipa Ridge block and the depth of incision into a pre-orogenic surface present on the San Gorgonio and Big Bear blocks.

The thermal history of Yucaipa Ridge block, constrained by apatite and titanite (U-Th)/He thermochronometric ages, estimates between one and two kilometres denudation must have occurred over the last ~1.25 Ma, assuming a thermal gradient of 30°C/km (Spotila et al., 2001). These values imply a long-term rate of denudation for the Yucaipa Ridge block of between 0.8 -1.6 mm/a (Table 7.1).
The weathered granitic horizon found on the Big Bear and San Gorgonio blocks (section 7.2) provides a marker horizon which has experienced minimal denudation since formation of the San Bernardino Mountains. By mapping the elevations of the weathered horizon remnants on the Big Bear and San Gorgonio blocks, Spotila et al. (1998) were able to reconstruct what the surface topography of the central and northern San Bernardino Mountains would have been prior to their formation. Assuming the initiation of orogenesis occurred 2.5 Ma ago (section 5.3.3), they subtracted the present day topographic surface from the reconstructed surface and derived the vertical rates of incision (section 5.6.1). Combined with the (U-Th)/He thermochronometric data they produced a map of long-term denudation rates for the San Bernardino Mountains (Fig. 7.5). Here, these rates are integrated over each basin, which has been sampled using cosmogenic radionuclide analysis, to allow direct comparison between denudation rates averaged over the intermediate-term with those over the long-term.

![Figure 7.5](image_url)

Figure 7.5 The map of denudation rates for the San Bernardino Mountains, adapted from Spotila et al. (2002). These rates were calculated based on incision into a pre-uplift marker horizon and from (U-Th)/He thermochronology.
7.3.3 Denudation rates over the short-term:

Sediment gauges monitoring the modern yields from the San Bernardino Mountains are not suitably placed to allow comparison of modern rates with the long and intermediate-term rates recorded from the small (0.3-8.4 km$^2$) basins sampled here. However, using the volumes of sediment measured since the 1940s by the Los Angeles County Department of Public Works, from a network of 75 small (0.5 to 8 km$^2$) debris basins along the southern slopes of the adjacent San Gabriel Mountains, Lavé and Burbank (2004) derive an average short-term denudation rate of 1.6 mm/a. The southern San Gabriel range front has a similar lithology to the Yucaipa Ridge block, it is predominantly gneiss and granitic, with a similar mean annual precipitation (Spotila et al., 2002) and mean basin slope (~31° from 30 m United States Geological Survey digital elevation model). Furthermore, with the rate of vertical displacement along the southern San Gabriel range bounding faults estimated to be ~1.0 mm/a during the Holocene (Blythe et al., 2002), it has comparable relative crustal uplift rates to those observed here for the Yucaipa Ridge block (Chapter 6). Hence, to obtain an estimate of short-term denudation rates, the southern San Gabriel range front is assumed to be analogous to the Yucaipa Ridge block in the San Bernardino Mountains.

7.4 Results:

At both the long and intermediate timescales, high rates are recorded on the Yucaipa Ridge block with more moderate rates recorded on the San Gorgonio block and the southern escarpment of the Big Bear block. These rates further decrease on the plateau and northern escarpment (Table 7.1, Fig. 7.6). The denudation rates in both
Figure 7.6 Denudation recorded over intermediate and long-timescales plotted against each other show a relationship where intermediate rates are approximately twice that of long-rates. However, there is significant scatter associated with the results due mostly to the stochastic nature of landsliding in rapidly denuding terrain.

data sets vary over three orders of magnitude. When plotted against each other there appears to be a ~two-fold increase in rates at the intermediate timespan in comparison to the long-term. However, there is significant variance of the denudation rates in landsliding areas due to the stochastic nature of the landsliding process in steep basins (Densmore et al., 1998, section 6.4.1). Furthermore, the long-term denudation rate of the entire Yucaipa Ridge block is derived from a (U/Th)/He thermal history and as such there is no appreciation of the co-variance of denudation rates from this block over the different timescales, although none would be expected unless the block had undergone significant tilting or internal deformation. Accordingly, mean rates for both the long and the intermediate-term, weighted by basin area, are calculated for each crustal block. The Big Bear block is further subdivided into three provinces based on geography. These are the plateau,
and the north and south escarpments (Table 7.1). Averaging the denudation rates in this way should give a macroscale appreciation of denudation from each province, provided several basins are used to derive the mean (section 4.5). Direct comparison of the mean rates averaged over the long-term with those derived using cosmogenic radionuclide analysis over intermediate timescales reveals a strong similarity as to where rates of rapid and relatively slow denudation are found. At an order of magnitude resolution, the denudation rates measured over the long and intermediate-term agree and show the same decreasing trend in denudation with distance from the San Andreas Fault Zone (Fig. 7.7). To illustrate this, the averaged denudation rates are shown for each block in the San Bernardino Mountains, including the short-term estimates derived from the debris basins study of Lavé and Burbank (2004) (Fig. 7.8).

7.5 Patterns of denudation:

7.5.1 Significance of differences:

The maximum difference between denudation rates, averaged over the long and intermediate-term, for each of the areas indicated in Figure 7.8, is ~four-fold. This is much less than the ~19-fold difference recorded over similar timeframes to infer increasing denudation in the Appalachian Mountains (Mills, 2000), but not dissimilar to the degree of variation recorded by Matmon et al. (2003) to infer temporal homogeneity of Appalachian denudation. Whether or not such variation constitutes a significant difference or implies uniformity of rates is, to some degree, subjective and so will be discussed further in section 8.3.2. The denudation rate differences measured here, across the north-south transect through the central San
Bernardino Mountains, records rates which vary over three orders of magnitude along a distance of a few tens of kilometres. Such a steep denudational gradient would be significant in most environments and, based on an examination of the factors maintaining denudation in the San Bernardino Mountains (Chapter 6), implies the San Andreas Fault imparts a significant influence on these rates, which lessens with distance from the main trace of the fault. Of particular note in this study, however, is that this denudational gradient appears at two different timescales.

Figure 7.7 Log-log plot of the mean denudation rates of the San Gorgonio block (SGB), Yucaipa Ridge block (YRB). The Big Bear block (BBB) is subdivided, based on geographical properties, into three provinces the northern and southern escarpments and plateau surface. Errors are the weighted means of the errors derived for each sample. For discussion as to how rates are derived see text.
Figure 7.8  Area-weighted denudation rates from the Yucaipa Ridge (for errors see table 7.1), San Gorgonio and Big Bear blocks with the Big Bear block separated into the northern and southern escarpments and plateau. Values in bold are long-term rates (from thermochronometric techniques or incision into the marker horizon), in bold with parentheses are intermediate rates (from cosmogenic radionuclide analysis) and bold in square brackets are short term rates (from sediment yield studies in San Gabriel Mountains (Lavé and Burbank 2004)). See text for discussion of different techniques.
and it is this general agreement, of this decreasing denudation rate pattern at the macroscale, on which the interpretation of consistency of rates over time is based. The argument of consistency of rates is made more convincing by the agreement of the short-term denudation rates from the nearby San Gabriel Mountains, the concordance of which implies that the denudation rates of the southern San Bernardino Mountains range front have been uniform over the last ~1.25 Ma.

The San Bernardino Mountains, therefore, show a temporal homogeneity or uniformity of denudation rates since initiation of their uplift. Phillips (2003) highlights the potential to incorrectly infer long-term consistency of denudation rates as sediment can be stored within a basin and evacuated incrementally, creating a buffered record of denudation which masks low-frequency high-magnitude events. Similarly, Lavé and Burbank (2004) record a temporal decoupling between sediment production and evacuation from basins >2 km² in the San Gabriel Mountains. However, the basins sampled in this study, except for those on the plateau, are too small and steep to retain much sediment. Cosmogenic radionuclide analysis should not be influenced by sediment storage of less than several hundred thousand years (Bierman and Steig, 1996) as it is recording radionuclide concentrations produced prior to detachment and transport. Furthermore, the thermochronometric technique is not influenced by storage of sediment and so the agreement between it and the cosmogenic technique suggests that storage issues are not the cause of the uniformity of rates.
7.5.2 Timescales of denudation:

Assuming then that the denudation rates recorded here are sufficiently consistent over differing timescales to warrant investigation, and that the consistency does not result from sediment storage buffering the rates, two further lines of investigation are considered: 1. what phenomena are the different techniques measuring and 2. what is the reason for the consistency of these rates over time? These points will be discussed in turn below.

Both thermochronometric and cosmogenic techniques are recording the rate at which a depth of crust is removed from the earth’s surface. In the case of thermochronometry, the temperature at which the He daughter product of radioactive decay is retained within mineral grains is assumed. Measuring the amount of He in grains at the surface allows an estimation of the time it has taken to remove a thickness of overlying bedrock with an inferred thermal gradient (Warnock et al., 1997; Wolf et al., 1998; Zeitler et al., 1987). Cosmogenic radionuclide analysis records the accumulation of radionuclides, produced by cosmic rays within a mineral lattice, as the depth of overlying bedrock decreases, and from this the rate at which a surface is denuding can be derived (Lal and Arnold, 1985; Lal, 1991). The depths of bedrock which must be removed before samples will have reached the surface is one to two kilometres and two to three metres for the thermochronometric and cosmogenic techniques, respectively, resulting in several orders of magnitude differences between the averaging times of the two techniques. Estimates of denudation from thermochronometric techniques provide a baseline rate for landscape development over the long-term. The long-term, million year time-frame
can incorporate several glacial/interglacial cycles and a multitude of individual seismic events such that any influence of these factors on denudation rates would be smoothed. By measuring denudation of only the top few metres of bedrock the cosmogenic technique is sensitive to denudational processes, to high-magnitude low-frequency events, and to recent environmental changes not recorded by thermochronometry. Hence, that the denudation rates for Yucaipa Ridge are similar when describing the removal of metres or kilometres of bedrock requires further consideration.

The rate of long-term denudation of the Yucaipa Ridge block has been derived from a model of thermal history based on a vertical elevation-age profile (Spotila et al., 2001). The Yucaipa Ridge block is a discrete block, too small to experience much internal deformation, so the rate of denudation derived from the removal of several kilometres of overlying bedrock would be consistent across the block unless it had been significantly tilted. However, it would not be expected that several metres of bedrock would be removed uniformly from the surface of the block over the intermediate or short-term, especially because the slopes are experiencing rapid mass wasting processes (section 6.4.1). The intermediate-term measurement is based on an average of six results which display a ~10-fold variation; the short-term rate is a mean of seventy-five results which display a ~60-fold variation (Lavé and Burbank, 2004). Matmon et al. (2003) suggest the variance of denudation rate decreases with increasing basin size because mixing from a population of sub-basins will begin to approximate a mean value. The results here quantify a decrease in denudation rate variance with increasing averaging time and suggests that, in

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tectonically active areas, the average of multiple short or intermediate-term rates from small basins can approximate the long-term rate.

What does this then say about the mechanisms driving denudation rates? Phillips (2003) points to an overwhelming tectonic influence to explain uniformity of denudation rates with respect to other environmental variables such as climate. This interpretation agrees with the lack of external environmental influence shown in Chapter 6, suggesting denudation at the macroscale in the San Bernardino Mountains is related to the rate of crustal input. The way in which tectonic mechanisms must be maintaining denudation is through base-level drop and fluvial incision as crust moves vertically across faults, as on the plateau escarpments, or by steepening slopes beyond the threshold limit of their mechanical strength, as on Yucaipa Ridge (Burbank 2002, Chapter 6). The temporal uniformity of denudation rates presented here from the Yucaipa Ridge block are averaged from several measurements which display increasing variance over shorter temporal scales. Intermediate or short-term fluctuations of rates could be the result of changing environmental factors. However, considering the variance of intermediate-term rates is greatest in areas of the San Bernardino Mountains where rapid mass wasting processes are prevalent, a stochastic denudational process related to the crossing of internal thresholds is more appropriate to explain the spread of rates. With slopes at their angle of repose, continuous slope failure occurring stochastically across the block characterises the Yucaipa Ridge. Because the average of the denudation rates from several basins over short and intermediate timescales is similar to the rate averaged over the last ~1.25 Ma, a mechanism able to maintain the consistency of
7.5.3 Implications of temporal homogeneity:

The similarity in the pattern of denudation rates over the different timescales suggests that the loci of relatively fast and slow rates have not moved over the lifespan of the San Bernardino Mountains and that there is a time-transgressive landscape response with distance from the San Andreas Fault. Meigs et al. (1999) and Lavé and Burbank (2004) have recorded consistent denudation rates over time in both the Santa Monica and San Gabriel Mountains (Fig. 7.1a), respectively, indicating temporally uniform rates of denudation rates may be common to the Transverse Ranges. The spatial pattern and temporal consistency of denudation rates show that crustal shortening has been principally focused on the main trace of the San Andreas Fault over the lifespan of the San Bernardino Mountains and is consistent with the notion of continued north-south compression across the eastern Transverse Ranges since the Plio-Pleistocene (Hill, 1982; Bird and Rosenstock, 1984; Humphreys and Weldon, 1986; Humphreys and Hagar, 1990). However, this would suggest that any changes in tectonic activity in the region which have occurred since the initiation of uplift of the San Bernardino Mountains (e.g. Morton
and Matti, 1993b; Matti and Morton, 1993; Hill, 1990) have not significantly
influenced the loci of crustal shortening. Furthermore, the north-south denudation
rate gradient does not appear to extend further west into the San Gabriel Mountains,
which display a more east-west oriented variation (Lavé and Burbank, 2004; Lifton
and Chase, 1992; Blythe et al., 2000). Does this imply the influence of the San
Andreas Fault is not uniform along the length of the Transverse Ranges through
which it runs? These issues will be discussed in more detail in the following chapter
which will expand upon many of the ideas raised in this chapter and in Chapters 3, 4
and 6.
8. General discussion

8.1 Introduction:

Several themes have been introduced in the previous chapters. Chapters 3 and 4 focused on the application of cosmogenic radionuclide analysis and what must be taken into account when applying the technique in the San Bernardino Mountains, southern California. The relative importance of the factors controlling the development of the San Bernardino Mountains was addressed in Chapter 6 and the measurement of denudation rate across a range of temporal scales in different part of the mountains in Chapters 6 and 7. In this chapter the discussions arising from this work will be expanded upon and discussed in relation to other ideas within a broader geomorphological context.

Firstly, the use of basin-wide cosmogenic radionuclide analysis in mountainous environments will be discussed, reviewing findings of Chapters 3 and 4 and discussing potential problems which arose from their application in terms of
sampling and the geomorphic processes being measured. The second section considers what constitutes a significant difference in cosmogenic denudation rate studies in relation to sources of error and will expand upon the findings of Chapter 6 in relation to concepts of thresholds, landscape sensitivity and mechanisms controlling denudation rates in the San Bernardino Mountains. Thirdly, the concepts of spatial and temporal scale, fundamental to any discussion of process rates, will be considered in light of the research in this study. Next, the implications of the results in relation to theories of orogenesis and the development of the San Bernardino Mountains will be presented. The last section will propose future areas of research for both development of the technique and investigation using cosmogenic radionuclide basin-wide analysis in tectonic geomorphological studies.

8.2 Application of cosmogenic radionuclides in mountainous terrain:

8.2.1 Validity of the basin-wide approach:

In studies of orogenic development the denudation rates sought using cosmogenic radionuclide analysis will often be rapid. When denudation is rapid, muogenic production provides a significant contribution to the total cosmogenic radionuclide inventory of the bedrock surface and so must be accounted for. The incorporation of muogenic production has the affect of increasing the mean attenuation length, and hence, the averaging time of the technique. Radionuclide decay must be ignored when employing the basin-wide approach to prevent error being introduced by spatially heterogeneous denudation (Bierman and Steig, 1996). In this study, the basin-wide approach of Bierman and Steig (1996) was considered with muon induced production included according to Granger et al. (2001). Results, plotted in
Figure 3.3, show that an increased mean attenuation length due to muons requires higher denudation rates if decay is to be ignored. If denudation is so slow that decay is significant and has to be included then an error may be introduced due to mixing of sediment from sub-basins of differing denudation rate. However, as denudation rates become lower the contribution of muons to the total cosmogenic radionuclide inventory decreases, reducing the mean attenuation length and thus the time available for cosmogenic radionuclides to decay before they are removed at the surface. This complex interplay of competing variables means a degree of error is introduced by employing the basin-wide approach, but that it can be minimised by validation of the appropriate assumptions. In rapidly denuding terrain, muogenic production is significant and rates of mass loss may be rapid enough that the affect of ignoring decay may have little impact on the denudation rate measured. In slowly denuding environments, decay becomes significant but ignoring the effects of a negligible amount of muogenic production will reduce the attenuation length which reduces the time available for radionuclide decay. Both scenarios allow application of the basin-wide model but the error introduced by mixing sediments from basins with heterogeneous denudation is minimised when denudation is rapid. In the $10^2$ to $10^3$ mm/a range of denudation rates a decision must be made as to whether 1. an overestimate of denudation rate will be introduced by ignoring decay, or 2. an underestimate incorporated by ignoring muogenic production, which will also provide an unknown amount of error resulting from potential heterogeneous denudation (section 3.3.4). Even with future developments in production rate accuracy, for basins experiencing this range of denudation rate, errors may always remain.
8.2.2 Sampling in mountainous terrain:
Collecting and interpreting cosmogenic radionuclide samples in mountainous terrain requires certain assumptions be considered explicitly, such as sufficient mixing, while others, such as sediment storage, are of less concern. Pulses of sediment delivered to channels are more likely to result in insufficient mixing in smaller basins because there is less time for alluvium to become homogenised (Sutherland et al., 2002; Miller and Benda, 2000). This can result in cosmogenic radionuclide concentrations that are not representative of the entire basin (section 4.5.2).
Furthermore, mass wasting processes have the potential to produce a grain size bias by unearthing larger, low concentration clasts from depth with respect to the high concentration, smaller size fraction removed from the surface (Brown et al., 1995). However, the need for large quartz masses in order to measure cosmogenic radionuclide concentrations in rapidly denuding environments has precluded more in-depth study of the phenomena in the landsliding environments where it is likely most significant. This has been principally due to the difficulties in processing sufficient quartz as it requires separating samples into smaller aliquots than might be practical to derive results with acceptable errors (Vance et al., 2003).

Sediment mixing problems can prevent application of basin-wide cosmogenic radionuclide analysis in non-landsliding environments also. Chapter 4 introduces a sediment mixing model, whereby, cosmogenic radionuclide ‘signatures’ may be identified and used to show the ratio of mixing at tributary junctions. This model has been applied by several others. Clapp et al. (2001) used it to investigate mixing in large low-relief environments. However, the radionuclide signatures from some
of the sub-basins they sampled are too similar to each other or the volumes of sediment leaving the smaller basins are insufficient in relation to the volumes from the larger basins, to reliably show the ratio of mixing between the basins. If upstream and downstream cosmogenic radionuclide signatures are similar, then mixing issues are of less importance as a source of error. Matmon et al. (2003) applied the model to show mixing was sufficiently thorough to ignore such affects on denudation rates derived from cosmogenic radionuclides in alluvial sediments of the Appalachian Mountains. In this case it was a reasonable assumption because of the two-fold difference between cosmogenic radionuclide concentrations from different sub-basins which provided discrimination between the source areas. As discussed in Chapter 4, differing cosmogenic radionuclide concentrations from different sub-basins within the San Bernardino Mountains are insufficiently mixed when derived from geomorphologically heterogeneous basins (section 4.5.2.2). The large errors associated with these samples, due to the rapid denudation rates, makes the degree to which samples are being mixed difficult to resolve. However, there exists the potential to use cosmogenic radionuclide signatures as a tool to investigate sediment mixing in fluvial geomorphological studies and to assess the impact of sediment mixing on aquatic biological habitats (Benda et al., 2004; Kirchner et al., 2001).

Most probably because of the potential for error introduced when measuring high denudation rates, there has been only one significant cosmogenic radionuclide study employing the basin-wide approach to estimate rates in a tectonically active environment (Vance et al., 2003). Other studies in such environments have avoided
the problems associated with calculating denudation using the basin-wide approach by sampling bedrock from strath terraces, estimating cosmogenic radionuclide exposure ages and from these a rate of fluvial incision (Burbank et al., 1996; Pratt-Sitaula et al., 2004). However, advances in the processing of cosmogenic radionuclide targets (e.g. Stone et al., in-press) and evidence that the technique can be successfully applied in areas of rapid denudation presents an opportunity to appreciate denudation from the microscale to the macroscale in tectonically active mountain ranges using analysis of cosmogenic radionuclides.

8.2.3 Denudational process

The early stages of orogenesis are often characterised by both steep slopes and uplifted low relief surfaces that are progressively removed as the slopes attain equilibrium with respect to base level change (Hovius et al., 1998; Schoenbohm et al., 2004). Application of cosmogenic radionuclide analysis throughout the San Bernardino Mountains has required sampling a variety of different terrains characterised by denudational processes, which range from the steep, mass-wasting, slopes of the mountains southern extents, to the transport-limited environment of the plateau. This section discusses the interpretation of the denudation rates measured in this study in relation to slope gradient and the geomorphic processes operating.

Previously studies have emphasised the relationship between basin relief and denudation (Pinet and Souriau, 1988; Ahnert, 1970, 1984; Schumm, 1963). However, the applicability of a relief-denudation rate relationship is not always appropriate (Montgomery and Brandon, 2002; Hovius, 2000). Relief normalised by
basin area or channel length may be a poor indicator of denudation at the small
Catchment scales ($10^6$-$10^1$ km$^2$) in this study because it does not cater for the
variable geometry of such basins (Hurtrez et al., 1999). One can envisage a
situation where mean basin slopes of two basins could be the same but their relief
ratio vary because of their shapes. This effect might be minimised by averaging the
data from many catchments but the thirty sampled in this study are unlikely to be
sufficient to achieve this. Furthermore, while calculations of the mean basin slope
may have previously been time-consuming the increasing availability of digital
elevation model data now makes rapid calculation of mean basin slope and other
geometric properties relatively simple (Burbank and Pinter, 1999)$^9$. Accordingly, in
Chapter 6, mean basin slope rather than relief was considered in relation to
denudation rate.

8.2.3.1 Threshold angles and denudational process:

Several studies note any relationship which might exist between slope and
denudation rate breaks down when slopes reach their limiting threshold angle, or
angle of repose; at which point they begin to fail by landsliding (Burbank et al.,
1996; Carson and Petley, 1970; Densmore et al., 1998; Montgomery and Brandon,
2002; Schmidt and Montgomery, 1995; Whipple and Tucker, 1999). The slope
angle at which this break, or decoupling, takes place has proven difficult to define

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$^9$ The mean slope calculated from digital elevation models may be sensitive to the cell size of the
data. In this study all mean slopes were derived from 10 metre digital elevation model data and this
resolution is considered to be fine enough to avoid significant dependency of the results on the grid
size of the data.
Figure 8.1 The relationship discussed in chapter 6 in which slope and denudation rate are related in an approximate linear fashion until 25-30° whereupon the relationship decouples.

with the maximum gradient of slopes being attributed to lithological strength (Schmidt and Montgomery 1995; Meigs et al., 1999), soil properties (Carson and Petley, 1970), climate (Gabet et al., 2004) and uplift rates (Montgomery 2001; Wobus et al., 2003). In the San Bernardino Mountains, the relationship between denudation rate and slope appears linear, or approximately so, until mean basin slopes of ~25-30°. Beyond this gradient range any increase in denudation rate is not matched by an increase in mean basin slope (Fig. 8.1). This indicates that at around 25-30° denudation rates decouple from slope in the San Bernardino Mountains and become a function of the base level drop at the foot of the slopes.

The lithology and climate in the landsliding portions of the San Bernardino Mountains are similar and preclude any interpretation of their influence on the threshold slope angles. However, the average slope gradient of Yucaipa Ridge and the southern San Gorgonio block is ~32°, in agreement with the threshold slope
angles obtained through digital elevation model analysis of the modal peaks of slope
distributions in the Himalayan and Olympic Mountains (Burbank et al., 1996;
Montgomery and Brandon, 2002). Appropriately, the basins in which the
relationship between slope and denudation has broken down correspond to those
undergoing landsliding and mass movement denudational processes. The point of
transition from fluvial incision to mass wasting processes has also been documented
in other parts of the Transverse Ranges to exist at ~25° (Campbell and Mackay,
1970) and to occur at ~0.3 to 0.4 mm/a (Lavé and Burbank, 2004). Hence, the near-
linear relationship between slope and denudation may allow prediction of rates in
other portions of the Transverse Ranges experiencing similar climatic and
lithological regimes, until slopes attain values in excess of ~25° and the denudation
rate becomes a function of incision at the toes of the slopes.

8.3 Climate, tectonics and the significance of difference:

Phillips (2002) considers the question ‘what constitutes a perturbation in the earth
sciences?’ How large does a change in environmental conditions, climatic, tectonic
or otherwise, have to be before it will affect the landscape? With measurements of
denudation it is possible to turn this question around and ask to what degree does the
landscape have to change before it can be considered to represent a change in
environmental factors? Does a difference in denudation rate over time or space
represent forcing by an external variable, or an internal threshold, and how can they
be differentiated?
Firstly, in this section the evidence for landscape change by climatic and tectonic forcing will be considered for the San Bernardino Mountains with explicit reference to geomorphic thresholds. Secondly, the degree to which measurements of denudation rate must differ over time and space before they become significant indicators of change is discussed in relation to cosmogenic radionuclide studies utilising the basin-wide approach.

8.3.1 Thresholds in the San Bernardino Mountains:
Evidence for episodes of fluvial incision or periods of landsliding are often used to indicate forcing by a climatic or tectonic mechanism (e.g. Schoenbohm et al., 2004; Snyder et al., 2000; Dadson et al., 2004). However, several issues complicate such a straightforward interpretation. Landscapes may exhibit complex responses to an initial perturbation such that they respond with fluctuating episodes of incision and deposition, which might be related to, but not directly driven by, a change in environmental circumstances (Schumm, 1973). Central to this concept are thresholds which may be crossed through external forcing or through the inherent internal dynamics of the geomorphic system (Schumm, 1979). Furthermore, the time taken for a landscape to respond to a change in environmental conditions may exceed the temporal scale of the environmental change, while some landscapes will be more sensitive to responding than others (Brunsden, 2001; Vandenbeghe, 2003). These factors hamper and often conflate determination of cause and effect when interpreting the landscape (Starkel, 2003).
Complex response is a natural oscillation within a geomorphic system as it attempts to return to an equilibrium state after a disturbance (Summerfield, 1991). This response reflects the linked nature of denudational, depositional and transport systems whereby a variation in one part of a system may resonate changes through the others. No matter how simple the system, the degree of change is unpredictable but maintained through the existence of feedbacks. These allow the breaching of internal thresholds and result in episodes of denudation and deposition which are driven by the system itself and not the results of any external changes (Humphrey and Heller, 1995).

Schumm (1979) categorises thresholds in landscape development as being either intrinsic or extrinsic to the system undergoing change. In the case of a colluvial or fluvial system, external thresholds may be crossed through variations in environmental factors such as climatic or tectonic change, whereas, internal thresholds are crossed through a change in variables within the system. Schumm (1979) further defines geomorphic thresholds as intrinsic thresholds integral to the development of a geomorphic system, such that landscapes may be inherently unstable and periods of denudation or deposition should result from a system's normal development. The presence of geomorphic thresholds means that evidence for accelerated or decelerated incision or deposition may not be evidence for a change in environmental variables but instead reflect a system attempting to equilibrate itself after the crossing of an internal threshold. For example, Densmore et al. (1997) note that the steepened toes of slopes can result from failure which is integral to slope development, rather than representing fluvial incision by trunk
streams which is often used as evidence for a change in environmental conditions. Tucker (2004) describes increases in denudation resulting from the crossing of an internal threshold for fluvial detachment. Humphrey and Heller (1995) model oscillations in sedimentation rates resulting from a complex response to an initial perturbation rather than a series of environmental changes. Identifying geomorphic thresholds requires a knowledge of the point at which these thresholds are likely to be breached, however, the lack of consensus as to the role played by climate and tectonics in active environments (Molnar, 2003) makes delineating geomorphic from extrinsic thresholds difficult. Indeed it is often defining the relative control of climate and tectonics on a landscape which is the goal of denudation rate studies.

So how might it be possible to identify whether changes in denudation rates either through space or time in the San Bernardino Mountains reflect changing environmental conditions or simply the crossing of geomorphic thresholds? There is a distinct spatial pattern of denudation rates which exists at the macroscale when the San Bernardino Mountains are split into provinces, on the basis of geography, and rates from the basins within those provinces are averaged (Fig. 8.2). However, this pattern is not as obvious when the rates are considered for each basin individually (Fig. 8.3). As shown in section 7.5.2, the rates derived using cosmogenic radionuclide analysis on Yucaipa Ridge, when averaged, are strikingly similar to the long and short-term denudation rates calculated using other techniques but individual basins which are adjacent to each other vary two-fold. There appears to be no lithologic, climatic or structural related explanation for this variation and so a logical conclusion is that it is the result of the development of steep slopes which are
Figure 8.2 The mean denudation rates derived from cosmogenic radionuclides along a transect in the San Bernardino Mountains grouped into five provinces based on the position of block bounding faults and geomorphology. The mean rates are the denudation rates from each basin weighted by that basin’s area.

Figure 8.3 Denudation rates recorded over intermediate and long-timescales plotted against each other show a relationship where intermediate rates are approximately twice that of long-rates. However, there is significant scatter associated with the results.
maintained through crustal input but which fail stochastically as internal thresholds are breached. While issues pertaining to sediment mixing (section 4.5.3) preclude confirmation that this is the case, the two-fold variation recorded on the landsliding portions of Yucaipa Ridge does not require a change in external conditions. It may be possible then to average out the influence of geomorphic thresholds at the small basin scale by using the mean rate of several basins (Schumm, 1975). However, this introduces the problem that all the basins draining into a trunk stream, Mill Creek in the case of northern Yucaipa Ridge slopes, may be influenced by the incision or aggradation of that trunk stream and it too is subject to geomorphic thresholds operating at a larger scale. To a certain degree the problem of misinterpreting change due to geomorphic thresholds as being the result of changing external variables has not influenced the conclusions drawn from San Bernardino Mountain denudation rate comparisons as they have tended to reveal a similarity rather than a difference. This similarity suggests either geomorphic thresholds are not significantly influencing rates in the San Bernardino Mountains, or that any influence thresholds have are smoothed by the spatial and temporal averaging of denudation rates over the different time scales. Where the variation between different timescales is noted on the Big Bear block plateau (section 6.4.4) it is so spatially consistent that it could not be the function of geomorphic thresholds as they would be breached randomly in each individual basin (Schumm, 1975). If all these basins on top the plateau drained into a single trunk stream such uniformity may be the result of the breaching of an internal threshold affecting the trunk stream which propagated up the drainage network into each of the smaller basins. However, these
basins do not drain into a single stream, hence, would not be influenced simultaneously as is occurring.

By considering the potential controls of denudation in individual basins and showing the consistency of the denudation rates from those basins it has been possible to infer the lack of influence of other environmental variables, such as precipitation, in comparison to tectonic controls (Chapter 6). Barring the notion that the consistency of rates on the plateau is coincidence, there is also a lack of geomorphic thresholds influencing the non-landsliding environments in the San Bernardino Mountains. However, slope failure in the steep regions of the San Bernardino Mountains such as on Yucaipa Ridge may be dependent on geomorphic thresholds effective over the $10^2$-$10^3$ year averaging times of the technique.

8.3.2 Significance of denudation rate differences:

8.3.2.1 Sources of error:

Watchman and Twidale (2002) point out that cosmogenic radionuclide methods provide numerical approximations and that while the empirical nature of the results is enticing, they are still approximations open to subjective correction factors. This has been shown in section 3.4.2.1 with the discussion of geomagnetic effects which change depending on the paleomagnetic record which is employed, and by the differing confidence in cosmogenic radionuclide production rates which vary from 5-20% for spallation of $^{10}$Be (Stone 1999; Dunai, 2003, pers. comm.). The more rapid the rate of denudation being measured the larger the error will tend to be. This is because more rapidly denuding samples will have lower cosmogenic radionuclide
concentrations, hence, lower isotopic ratios (e.g. $^{10}\text{Be}/^{9}\text{Be}$), to the point where they become indistinguishable from the isotopic ratio of the sample blank$^{10}$. This error can be minimised by reducing the amount of carrier (e.g. $^{9}\text{Be}$) added or increasing the sample mass in order to increase the amount of cosmogenic radionuclide (e.g. $^{10}\text{Be}$) being measured. For a denudation rate of 1 mm/a, assuming a 20% production rate uncertainty, an analytical AMS uncertainty of around 7%, a carrier mass of 350 μg and a quartz mass of 30 g, the propagated error is ~65%. The same values for a denudation rate of 0.1 mm/a produces a propagated error of ~10%. Clearly then the sample error in regions of lower denudation rate will be reduced in comparison to high denudation rate environments.

Much of Chapter 4 discussed the problem of sediment mixing and the error this might introduce. It is more problematic to estimate the error which might occur through insufficient mixing than it is to estimate the random error due to uncertainty in the variables used in the denudation rate model. Where insufficient mixing may be an issue, such as at channel junctions, it can be quantified using the method employed in section 4.5.2. When sampling low order streams, however, the lack of channel junctions prevents the comparison of sediment from different parts of the basin. One solution to this is to average several samples, as was done for Yucaipa Ridge in section 7.4, to obtain a mean value. This requires the assumption that all the basins being averaged have equable denudation rates. Whether such an average would be representative if basins had very different denudation rates is less certain.

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$^{10}$ Sample blanks are used to measure background Be concentrations introduced to the samples during target preparation. The Be concentrations measured by the blanks are subtracted to obtain cosmogenic Be only (section 3.6.2.2).
Hence, in mountainous environments where denudation rates are high and sediment mixing sporadic it is difficult to quantify a degree of uncertainty for samples. It could be argued that issues of sediment mixing preclude interpretations of the results. However, the agreement of the pattern of denudation rates averaged over different timescales at an order of magnitude resolution shows they are able to provide at least first-order estimates of the rates of surface processes in rapidly denuding terrain. To investigate this issue further the degree of difference used in current literature to infer a change, or lack of, will be considered.

8.3.2.2 Degree of difference:

The basin-wide approach has been applied in a range of environments in order to investigate spatial and temporal patterns of denudation (Chapter 2). However, what has either constituted change, or uniformity, of rates in these studies has varied. Several studies have concluded that change in denudation rates over time occurs, based on the comparison of modern sediment yields with cosmogenic radionuclide analysis. The degree of this change varies from a 1.5 to 4-fold increase (Schaller et al., 2001) and 1.5 to 10-fold increase (Schaller et al., 2002) in cosmogenic radionuclide derived denudation rates for large, predominantly lowland, basins. A <two-fold decrease (Clapp et al., 2000), a 17-fold mean increase (Kirchner et al., 2001) and a 10 to 100-fold decrease (Hewawasam et al., 2003) in cosmogenic radionuclide derived rates for smaller mountainous catchments are used to support change. Studies which have concluded a temporal homogeneity of rates include a <two-fold decrease from a small low-relief basin (Clapp et al., 2001) and a ~two-fold decrease from mid-sized (Matmon et al., 2003) and small (Brown et al., 1995)
mountainous basins. Studies comparing the cosmogenic radionuclide derived rates with long-term results have used a <four-fold increase (Matmon et al., 2003), a ~two-fold (Kirchner et al., 2001) and ~zero to five-fold decrease (Bierman and Caffee, 2001) in cosmogenic radionuclide derived denudation to infer uniform rates over longer timescales. Clearly there is some ambiguity as to what constitutes change and what uniformity, with different studies arguing the same degree of difference represents both temporal heterogeneity and homogeneity. While order of magnitude differences in denudation rates measured over different timescales by several studies can confidently be assumed to indicate a change has taken place, those studies where the differences are only a few-fold are less conclusive. The Clapp et al. (2000) and Schaller et al. (2001) argument for change is unconvincing based on the comparison of cosmogenic rates and modern sediment yields when the error associated with modern sediment yields, the uncertainty of cosmogenic radionuclide production rates, and the difficulty in verifying the assumptions inherent to cosmogenic radionuclide analysis is considered. In Chapter 6 the maximum difference between the long and intermediate-term denudation rates, taken from an average of several basins, displays ~4-fold higher cosmogenic radionuclide derived rates on the southern escarpment of the Big Bear block. While individually this result may be interpreted as representing change, when it is considered along with the rest of the denudation rate data from the San Bernardino Mountains this is the maximum difference between rates which vary over nearly three orders of magnitude. Hence, it implies that, at the macroscale, rates have shown little change over the long and intermediate timespan.
Generally, the above studies show much less variation in denudation rates when they are compared over long and intermediate timescales in comparison to the changes observed between the intermediate and the short-term. This would be expected as short-lived, high magnitude events, which occur only over brief timescales are smoothed over increasingly long periods (Wolman and Miller, 1960). Furthermore, the expected anthropogenically forced increase in denudation rates at shorter timescales (Douglas, 1967; Saunders and Young, 1983) appears inconclusive. The discussion in section 8.2.3.1, as to the dependence of cosmogenic radionuclide derived denudation rate error on the rate measured, means that for the lower denudation rates recorded in some of the above studies, smaller differences between rates may become more significant. For cosmogenic radionuclide studies in rapidly denuding terrain the uncertainties are still large and more obvious differences must be sought to represent change. Furthermore, the potential for geomorphic thresholds means that their role must be identified in a landscape system before any change is considered to result from forcing by external environmental variables (Viles and Goudie, 2003).

8.3.3 The roles of climate and tectonics:

Despite the issues arising from geomorphic thresholds it has been possible to identify a tectonic control of denudation rates in the southern San Bernardino Mountains. Heavy precipitation, along with seismic shaking, will trigger much of the mass wasting of the steep Yucaipa Ridge block slopes (Lavé and Burbank et al., 2004; Keefer, 1994; Minnich, 1989). However, to maintain their steepness over the long-term a base level drop at the foot of the slopes is required and is facilitated by
vertical crustal movement (Chapter 6). Because the Yucaipa Ridge block is bound
by major low-angle strands of the San Andreas Fault it is unable to widen and so
cannot increase in height (Ahnert, 1984). The constancy of the denudation rate over
the long, intermediate and short-term suggests this has been the case for the last
~1.25 Ma. If the rates measured here are indicative of slopes at their limiting
threshold angle, as is implied by Figure 8.1, then this block has maintained a
topographic and denudational steady state over this period, facilitated through base
level change at the toes of the slopes and a mass wasting response. If the relative
uplift of the San Bernardino Mountains does not predate three Ma ago, as is
indicated by provenance of local deposits (section 5.3.3), there is the implication that
thin rapidly denuded crustal slivers, such as the Yucaipa Ridge block (Blythe et al.,
2002; Spotila et al., 2001), are able to reach an equilibrium form between crustal
input and denudation in less than three million years. Although the input of crust
maintains the threshold slopes on Yucaipa Ridge this does not exclude the potential
for climate to exert influence. Gabet et al. (2004) suggests that increased
precipitation correlates with characteristic lower slopes in the Himalayas. If there
was an increase in the rate of precipitation in the southern San Bernardino
Mountains, the slopes of Yucaipa Ridge might attain a lower threshold angle of
repose. However, this would cause an increase in denudation rates only until the
new threshold slope angles were achieved, because once slopes are at their angle of
repose denudation rates become a function of base level drop not slope (Burbank et
al., 1996; 2002; Hovius et al., 2000; Densmore et al., 1998; this study Chapter 6).
There is no evidence precipitation exerts an influence on denudation rates in the
non-landsliding portions of the San Bernardino Mountains (section 6.4.2.1).
may be because mean annual precipitation, used here, is not related to geomorphic work and instead the seasonality of precipitation should be considered (Ahnert, 1970; Douglas, 1967; Venderberghen, 2003; Carson and Petley, 1970). It may also be because any climatic influence is dwarfed next to forcing by tectonic mechanisms (Phillips, 2003) or that climatic variability influences denudation on short timescales (Viles and Goudie, 2003) and this is not recognised by the $10^3$-$10^4$ year averaging time of the cosmogenic radionuclide derived rates. It does imply, however, that some studies are incorrect in assuming denudation rate is some function of mean annual precipitation.

8.4 Scales of measurement:

Throughout preceding discussions the importance of spatial and temporal scale has been stressed. Both when sampling and interpreting the results the basin size should be considered. Denudation rates derived at a particular spatial or temporal scale are recording phenomena which may not exist at different scales (Schumm and Lichty, 1965). This may hamper comparison or allow information to be gleaned but requires acknowledgement of the scales used. This section will further arguments made in previous chapters and discuss the issue of scale with respect to the work of others.

Firstly, basin-wide cosmogenic radionuclide analysis relies on certain assumptions that can be violated without prior consideration of the scale of the basin and the processes operating within it. Hence, alluvium sampling issues will be briefly considered before going on to present a discussion of the relevance of spatial and
temporal scales in the interpretation of basin-wide denudation rates over a variety of timespans.

8.4.1 Spatial scale:

8.4.1.1 Sampling:

The homogenising effect of sampling at an increasingly larger basin scale has been described by Matmon et al. (2003), such that, as the basin size increases the denudation rate variance recorded by cosmogenic radionuclide samples decreases. Increasing basin size will increase channel length and the time there is available for mixing to occur. In the small steep basins sampled in the San Bernardino Mountains there may be little time for the sediment to become thoroughly mixed. For samples collected from channels in steep sided gorges the occurrence of localised deposition into the channel from surrounding slopes may result in a sample not representative of the entire upstream area (section 4.5.1). For the larger basins sampled by studies such as Schaller et al. (2001, 2002), the increased channel lengths would provide homogenised alluvium and the lowland sample sites reduce the possibility of localised depositional events but as the basin scale becomes larger the issue of sediment storage becomes more problematic. The trade off between errors introduced through insufficient sediment mixing or sediment storage means there may be an appropriate basin size which minimises the problems of both. However, the basins sampled in the San Bernardino Mountains were small enough that sediment mixing was potentially problematic and accordingly it was considered in detail (Chapter 4).
Another issue pertinent to basin scale is the buffering effect which sediment storage might have on sediment volumes leaving a catchment. Provided sediment storage is not longer than several 100 kathe cosmogenic denudation rate results should not be affected as they rely on cosmogenic radionuclide concentrations produced prior to detachment from the surficial bedrock. However, when comparing cosmogenic radionuclide results with sediment yield data it must be remembered that sediment produced by high magnitude events may not be transported synchronously from the basin. In the case of landsliding terrain there is the potential for a decoupling between sediment derived from hillslopes and the transport of that sediment (Hovius et al., 2000). This may act to buffer or regulate the volumes of sediment transported from a basin (Phillips, 2003). Larger basins would be more susceptible to this effect (Lavé and Burbank, 2004). This could either help or hinder comparisons of cosmogenic radionuclide derived denudation with short-term studies as one technique is a function of the residence time of bedrock at the surface before detachment and the other a function of the residence time of that detached bedrock during transport within the basin. Differences between the two record the coupling between the detachment and transport of mass.

8.4.1.2 Interpretation at differing basin scales:
Assuming problems related to mixing of sediment are avoidable, or can be minimised, how then does scale influence the denudation rate measured by cosmogenic radionuclide analysis? The size of the basin sampled is considered to reflect the sensitivity of that basin to change (Brunsden, 1980; Vandenberghe, 2003). The small basins sampled here would be expected to respond rapidly in
comparison to larger ones. Results here indicate this is not always the case. If base level change at the foot of a slope migrates upstream as a knickpoint (Sugai and Ohmori, 1999) it may take some time for regions at higher elevations to experience the affects of the change. The high denudation rates at the base of slopes of the San Gorgonio block and low denudation rates at higher elevations, where remnants of pre-uplift topography remain, highlights this. If environmental change is communicated to a basin through fluvial incision of channels the higher, smaller parts of a basin may be the last to respond (Harvey, 2002). Furthermore, the ability for a basin to respond to change may depend on how close it is to a threshold. In the case of Yucaipa Ridge, rapidly denuding basins at threshold angles would require little to make them fail while the more slowly denuding, more gentle slopes of one basin on the ridge (Chapter 6; sample TC) would respond by more gradual fluvial incision in response to changing environmental conditions. These basins are the same approximate size but show different sensitivities because of their different propensities to respond to change. A complex interplay then exists between spatial scale and external forcing and while basin size may be broadly applicable as a measure of sensitivity it is the steep basins or the parts of basins closest to locations of base level change which are most sensitive. Schmidt and Montgomery (1995) discuss the relationship of scale to lithologic strength when considering the variables controlling denudation of mountains, acknowledging that small scale laboratory tests do not accurately estimate the strength of rock at the scale of mountains. Applying laws derived from small-scale studies to the broader landscape is problematic because thresholds exist at different spatial scales (Harvey, 2002; Lawrence, 1996; Martin and Church, 1997; Tucker, 2004). Furthermore, changing the spatial scale of
measurement may change the mechanisms that are driving the rates of denudation (Schumm and Lichty, 1965). For example, Arnett (1979) suggests a reduced influence of climate on denudation rates as basin scale decreases. Hence, measurements from small basins along a transect through the San Bernardino Mountains may not be pertinent to macroscale issues of orogenic development. Whether or not averaging the denudation of several small basins in a region can be considered to represent the denudation of that region when different thresholds might exists at larger spatial scales is questionable. In certain cases in the San Bernardino Mountains there is no option but to use the average of several small basins to represent a larger area, such as on Yucaipa Ridge block, where channel reaches are short and basins do not get larger than those sampled in this study. In some parts of the San Bernardino Mountains, variable lithologies or significant anthropogenic impact prevented the sampling of larger basins. However, denudation occurring due to thresholds being breached at a large scale, for example a global fall in sea-level, would propagate up the drainage system as knickpoints till it affected smaller scales. The question then would concern the time it took for this to happen and whether or not samples were collected upstream and downstream of the knickpoint and subsequently averaged (Harvey, 2002).

There are clearly issues to do with the application of the cosmogenic technique in basins of different sizes resulting from sediment mixing. Basin size might also reflect on sediment storage, influencing rates derived by sediment yield studies. Sensitivity to environmental change and thresholds are a function of basin scale and may make it difficult to extrapolate the results from one scale to another. However,
this could also be utilised to help identify the scales at which thresholds operate. The spatial scales of processes are linked closely to their temporal scale through notions of frequency and magnitude and so it is time scales which are considered next.

8.4.2 Temporal scale:

Denudation rates derived using cosmogenic radionuclide concentrations embody an averaging time characteristic of the different production rate mechanisms and the rate of denudation being measured (section 3.5). This averaging time is exponentially weighted towards the present as the rate of cosmogenic radionuclide production increases exponentially towards the surface. This means the majority of the cosmogenic radionuclide concentration, from which a rate of denudation is derived, is bias towards the present; although this effect has not been empirically incorporated in the weighted mean averaging times derived for the San Bernardino Mountains samples.

One assumption of the cosmogenic radionuclide denudation rate model is that rates have been uniform over a period of time long enough for cosmogenic radionuclide concentrations in bedrock to achieve secular equilibrium with respect to production, loss and decay (Lal, 1991). The results from the San Bernardino Mountains support, at the broad-scale, the notion that rates can be consistent enough to validate this assumption, even in rapidly denuding mountainous environments. However, whether or not this has been the case is difficult to test for individual samples. $^{10}$Be/$^{26}$Al ratios can be employed to investigate assumptions of steady state
denudation utilising the isotopes differing decay times (Lal, 1991). However, the large error associated with $^{26}$Al results (Fifield, 2003, pers. comm.) and the diminished influence of decay in rapidly denuding terrain (section 3.3.2) prevents this application in the San Bernardino Mountains.

Assuming that methodological pitfalls have been avoided, how might the averaging time of a denudation rate measurement influence the rate measured? Wolman and Miller (1960) presented the case for the frequency and magnitude of events and concluded that moderate sized events do most geomorphic work in fluvial systems. Hence, low-magnitude events may happen most often but their cumulative effect is small and while high-magnitude events have catastrophic impact they occur too rarely over the long-term. However, work on the scaling of landslides has indicated that large, low-frequency events are more important in terms of the amount of material they produce (Hovius et al., 1997; 2000). In a separate study Reid and Page (2002) concluded that moderate events with a recurrence interval of less than 27 years were responsible for the majority of sediment delivered to alluvial channels in landsliding environments. The results from the southern San Bernardino Mountains would support the notion that such moderate sized landsliding events were producing most sediment as the rates over decades from a region analogous to Yucaipa Ridge match those over longer timescales. Were large, low-frequency events producing the most sediment, the intermediate or longer-term rates would be higher than the shorter. However, as mentioned previously, while climatic or seismic events might trigger landsliding on Yucaipa Ridge it is continuous failure due to crustal input that steepens slopes beyond their threshold that must be the
underlying mechanism behind the denudation rate. Because of this and possibly aided by the fractured nature of the bedrock, high-magnitude, low-frequency events are not observed as it takes very little to make these slopes fail. As such, Yucaipa Ridge slopes, which are constantly unstable, may provide accurate estimates of long-term denudation over decadal timescales provided any spatial variation in rates due to stochastic failure is accounted for (section 8.3.1). While landslide scaling laws may be derived for individual locations there is unlikely to be one rule to fit all and frequency magnitude effects might be subject to localised conditions.

In their classic 1965 paper, Schumm and Lichty proposed that temporal and spatial scale might control which mechanisms drive landform development. For example, at the broadest scales tectonic influences would dominate whilst at smaller scales processes would be independent of tectonic forcing and instead variations of factors such as hydrology would be causal mechanisms of change. Harvey (2002) considered scale and the coupling between different components of the landscape and concluded that as scale changes so do controlling factors but that localised changes occur within the framework of the broader scale. As such, climatic factors may control denudation at the scale of hillslopes but will be operating within the context of any base level changes. The notion that one factor may overwhelm others in driving rates of denudation is proposed by Phillips (2003) to explain the consistency in rates of denudation measured over different timescales. The consistency in rates over differing timescales in the southern San Bernardino Mountains may well be a function of a tectonic influence dominating any climatic or more localised influences. However, in non-landsiding localities within the
mountains to which knickpoint migrations have yet to carry the signal of crustal uplift and base level change there appears to be little influence of other factors. This indicates that either climate does not directly affect denudation rates (Riebe et al., 2001) or at least that mean annual precipitation is inadequate to represent any influence (Ahnert, 1970). Alternatively it could be argued that climate does not affect denudation rates at the scale of small basins measured or at the $10^3$ to $10^4$ year timescales of the cosmogenic radionuclide technique in the San Bernardino Mountains, but that it has a more regional or short term impact.

8.5 Implications for the evolution of the San Bernardino Mountains:

Geomorphological models of landform development have progressed from qualitative interpretations (Tinkler, 1985), to those which incorporate more empirical contributions and recognise the Earth’s surface as an interface between competing tectonic and climatic forces (e.g. Hovius, 2000). Increasingly, computer simulation is used to contrast and compare the interactions of these forces (Willett et al., 1993; Kooi and Beaumont, 1996). To circumvent problems of quantifying rates of denudation and surface uplift in tectonic environments many models assume steady state conditions exist, whereby crustal input is balanced by mass removal. While such studies are more often applied at a larger spatial scale than is provided by the San Bernardino Mountains, the wealth of empirical data for this range provokes a discussion of the existence of steady state within the San Bernardino Mountains. Secondly, the implications of the results presented here for the orogenesis of the San Bernardino Mountains, within the framework of their
evolution discussed in Chapter 5, is presented. Lastly, the results of this study will be considered in a more regional context.

8.5.1 The San Bernardino Mountains and theories of orogenesis:

Hovius et al. (1997) point to the influence which large scale drainage distribution has in ordering topography. Headward expansion of basins via landsliding cuts into nascent highlands causing stream capture and drainage reorganisation until intermontane basins become sufficiently mature and the fluvial network sufficiently entrenched that major events are required to change the form of the landscape. In this way, uplifted topography tends towards a steady state with respect to base level. Ahnert (1970; 1984) presented a similar simulation when discussing how the control of the height of mountains is the width of the range. Everything else being equal, as slopes tend towards their threshold angles wider ridges will become higher\(^{11}\). The San Bernardino Mountains are a young range which appear to exhibit both these traits.

The large-scale drainage patterns in the San Bernardino Mountains follow the surficial trace of the major block bounding faults. The Santa Ana Valley and Mill Creek intermontane basins are structurally constrained channels not fluvially incised features. The location of the North Frontal Thrust Zone and the Santa Ana Thrust Fault, and the Mill Creek and San Bernardino Strand Faults, defines the width of the crustal blocks comprising the San Bernardino Mountains and thus constrains the

\(^{11}\) Until they become too high to support their own weight and undergo widening through gravity collapse.
height and slope angles of the blocks. The broad Big Bear block exhibits the lowest mean slope angles and the thin Yucaipa Ridge block the greatest. Dibblee (1982) described the plateau of the Big Bear block as an uplifted erosion surface, that is to say a surface which has been denuded until it has little relief. Although stated in a context of Davisian landscape evolution it is a suggestion now backed by much evidence indicating the plateau surface has maintained low relief throughout the uplift of the San Bernardino Mountains (section 5.5.1.2). It is further supported by this study which derives low rates of denudation for the plateau. Assuming the plateau is a surface which has been altered little since uplift it presents a location for the testing of ideas concerning the initial stages of orogenic evolution. Rapid denudation along the southern escarpment has given way to stream capture of part of the plateau drainage network through Bear Creek at ~1.5 Ma (Cox et al., 2003). The most rapid rates of denudation along the southern escarpment are found in basins on either side of where Bear Creek has incised a steep gorge. That this is also where the escarpment exhibits the greatest relief may not be coincidence, but instead supports the notion that the most rapidly denuding parts of the escarpment are defining the location of future drainage as the plateau becomes more incised. Spotila et al. (2002) describe the plateau surface as incising to achieve a rougher topography such that over time it will come to resemble the form of the older San Gabriel Mountains. The evaluation of denudation rates over different timescales provides evidence that the rate of denudation over the last $10^4$ years is approximately twice the rate averaged over the ~2.5 Ma lifetime of the mountains. This provides a timeline for the development of the plateau and suggests that the pace of incision by
upstream knickpoint migration due to base level change around the perimeter of the block and subsequent slope steepening is too subtle in comparison to the rate at which stream capture events like the Bear Creek example have altered the form of the plateau. The topography of the Big Bear block is thus at a crucial stage of development where the future form and height of the range will be decided by the emplacement of a new drainage network.

Because of the width of the Big Bear block the plateau surface has survived and slopes have low gradients. The Yucaipa Ridge block, however, is thin and being denuded rapidly through base level drop along the toes of its slopes. The uniformity of the rates of denudation over different timescales on Yucaipa Ridge implies that the rates are time independent, defined by Burbank (2002) as an ‘erosional’ steady state. Furthermore, as any crustal uplift of the block must be matched by at least the same amount of mass loss to maintain slopes at their threshold for failure (Burbank et al., 1996), the block is in a state of equilibrium with respect to the flux of crustal input and the rate of denudation. This condition is defined as ‘topographic’ steady state (Burbank et al., 2002). Steady-state concepts are more often used to describe entire mountain ranges rather than a single crest as in the case of Yucaipa Ridge. However, although the concept of topographic steady state is used to circumvent problems relating to the quantification of uplift and denudation in numerical and computer landscape simulation models (Snyder et al., 2000), there remain relatively few conclusively documented cases where rates of processes are known (Ducea et al., 2003). Yucaipa Ridge then would appear to offer a natural laboratory to refine model algorithms. Bedrock channel evolution in landscape evolution models is
usually envisaged as the resistance of bedrock shear stress on the channel bed, derived using the drainage area as a proxy for precipitation and channel slope (Howard and Kerby, 1983). This study proposes that, in tectonically active orogens, slope may be the more important variable and that mean annual precipitation has little influence. Furthermore, the denudation rate results presented here suggest for landscape models incorporating slopes steepened above 25-30° stochastic slope failure, as discussed by Densmore et al. (1998), becomes a fundamental component.

The San Bernardino Mountains exhibit a transition from time independent to time dependent landscapes with distance from the trace of the San Andreas Fault. This appears in part to have depended on the dimensions of the blocks and the distance between bounding faults. Whether proximity to the San Andreas Fault has helped promote this or whether the width of the crustal blocks making up the mountains would be expected to increase in size with distance from the fault may help further constrain models of landscape evolution and orogenic development. It is this broader context which is considered next.

8.5.2 Regional context:
The San Bernardino Mountains have risen as a result of compression across the Big-Bend of the San Andreas Fault Zone. Denudation rates through time in the San Bernardino Mountains help constrain orogenesis in an active fault zone. The above discussions have implicated tectonic mechanisms for the pattern of decreasing denudation rates with distance from the San Andreas Fault. Assuming this to be the
case and that any climatic influence is minimal, then there is a relationship between the rates of denudation and the distance to the trace of the main fault zone. At the mesoscale, Riebe et al. (2000) suggested that the proximity of slopes to a fault or incising channel is a control of the denudation rates in the Sierra Nevada (section 2.3.2). Here the broad trend of denudation at the macroscale is also one of decreasing rates with distance for the main trace of the fault and displays a decline in rates by two orders of magnitude over spatial scales of a few tens of kilometres. This is evidence that the influence of major faults on landscape evolution can be limited to a narrow zone along their length. Whether the influence of the San Andreas Fault Zone operates in this manner along the length of the fault through the Transverse Ranges seems less feasible though. Estimates of denudation in the San Gabriel Mountains show an east to west decrease along the trace of the San Andreas Fault rather the south to north decrease away from the fault found in the San Bernardino Mountains (Lavé and Burbank, 2004). This is in agreement with the east to west decrease in tectonic activity inferred from the stream gradient index and fault activity in the mountains (Keller, 1986; Lifton and Chase, 1992) further supporting the notion that tectonic mechanisms control the macroscale landform development of the Transverse Ranges. If the form of topography is reliant on the dimensions of crustal blocks as suggested in section 8.5.2 then the flake tectonics model, whereby fault zones comprise discrete crustal blocks, has ramifications for the denudation rates recorded in active mountain ranges.

The variability in denudation rate patterns between the San Bernardino and San Gabriel Mountains highlights their individual development. However, in each of the
San Bernardino Mountain, San Gabriel Mountain and Santa Monica Mountain localities in the Transverse Ranges, uniform rates of denudation through time have been recorded (Burbank and Lavé, 2004; Meigs et al., 1999; this study). If movement on the San Andreas Fault is driving the broad patterns of denudation throughout the whole Transverse Ranges then the uniform influence of the fault over time could be suggested as the reason for the consistency of rates. However, much geological evidence points to the changing nature of the location, rate and style of faulting in the San Andreas Fault Zone over the last several million years (Blythe et al., 2002; Cox et al., 2003; Matti and Morton, 1993; Morton and Matti, 1993b; Powell et al., 1993). Gunnell (1998) postulates that consistency of denudation rates in the cratonic Dharwar region of the south Indian shield results from a slope system that has not been subject to significant modifications. It seems more feasible that this is the reason for constant rates over time in the San Bernardino Mountains.

Because the major drainages in the San Bernardino Mountains, the Mill Creek and Santa Ana River, have had to negotiate the relative uplift of crustal blocks, they have been forced to follow the trace of the block bounding faults. As such, significant drainage reorganisation has been prevented from occurring in the mountains except on the plateau. That is to say the location of ridges, escarpments and plateau in the San Bernardino Mountains have not yet changed significantly since uplift began, and this has constrained the large scale drainage pattern in the mountains. This pattern will change, however, as the topography of the Big Bear block evolves and stream capture and drainage reorganisation ensues. In the San Gabriel and Santa Monica Mountains the same explanation may apply, or it may be that in these older ranges topography is mature enough to be little modified because drainage spacing
and the subsequent tempo of landscape response has existed sufficiently for long and short-term rates to agree.

8.6 Future considerations:

From the discussion above it is clear that several areas of understanding are lacking both in terms of the application of cosmogenic radionuclide analysis in high relief environments and in the understanding of the effects of temporal and spatial scale on the results obtained. These themes are considered below along with potential areas for future research within tectonic geomorphology.

With the exception of Vance et al. (2003), studies have not applied cosmogenic radionuclides to study rapidly denuding, basin-wide process rates in tectonically active environments. As the potential to reveal fundamental relationships behind the processes which shape active margins is vast this is likely to change. However, several areas of the technique will require further study before it becomes more widely applicable. These include the confidence in the technique in areas of active landsliding. The results above suggest areas of shallow landslides are accessible to the technique but require a differing set of assumptions than might be used in a low relief environment. Perhaps one solution to the problem of adequate sediment mixing in landsliding environments lies in computer modelling. Rather than attempting to obtain rates of denudation using radionuclides, the cosmogenic radionuclide inventory of a landscape undergoing rapid mass movement could be simulated and tested via measurements of concentrations found in the bedrock, colluvium and alluvial sediment of a landsliding environment (Burbank, 2004, pers
comm.). In this way assumptions of how sediment mixing might invalidate the technique could be tested. Such study would likely have to focus on patterns of cosmogenic radionuclides across a landscape rather than absolute values of denudation but would allow insights into the transport and mixing of mass in high relief environments. Another problematic issue when applying cosmogenic radionuclides in highland regions is the presence of glaciation. If glacial limits are well known the shielding effect ice will have can be estimated. However, the depth of the ice may be harder to define and if muogenic production is included it may be that neutron spallation production in sub-glacial bedrock is prevented but muogenic production is not. These problems must be dealt with before the technique is readily applicable to high, mountainous regions.

Another aspect of cosmogenic radionuclide analysis that must be refined in order for it to become more applicable is production rate. The ongoing CRONUS project is an international effort that seeks to refine estimates of production and methods of scaling for cosmogenic radionuclides. This may bring the uncertainty on production rates closer to the more acceptable levels of analytical uncertainty associated with accelerator mass spectrometer measurements, allowing greater resolution of calculated denudation rates. However, such an advance must be matched by a better understanding of sediment mixing for the benefits to be gained by the basin-wide approach in mountainous areas.

The most surprising results of this study are that climatic variation, across the range of environments addressed in this study, exerts no significant influence on the rates
of denudation measured and that the pattern of rates measured in the San Bernardino
Mountains are the same over varying timescales. Molnar (2003) calls for a return to
theory to order the respective roles of climate and tectonics in geomorphology.
While a sounder theoretical framework on which to base testing would be beneficial,
testing via accurate estimates of denudation rates at a variety of scales will still be a
necessity. To better understand the influence of climate it would be prudent to select
regions where the effects of crustal uplift climate and lithology can be isolated.
When several of these areas, under differing climatic regimes, had patterns of
denudation rates derived using cosmogenic radionuclide studies, the role of climate
would become more apparent. Undertaking the denudation rate measurements at a
range of temporal and spatial scales would allow the issue of the scales over which
climate may be an influence to be addressed.

Finally, there is a growing number of global positioning system arrays in California
and other tectonically active regions that are beginning to elucidate the motion of the
crust in fault zones. At present, networks in the Transverse Ranges area such as the
Southern Californian Integrated Global positioning system Network\textsuperscript{12} are limited in
extent to macroscale crustal movement and have only been collecting data for
several years. However, as the distribution of array stations becomes denser, and
has been operating longer, the short-term crustal deformation will become apparent.
Comparison of crustal movement and denudation could facilitate advances in the
understanding of denudational response to crustal motion and the feedback

\textsuperscript{12} See http://www.scign.org
mechanisms which may exist between these two processes (Willett et al., 1993; Koons, 1989).
9. Conclusions

The work presented here combines the testing of the capability of basin-wide cosmogenic radionuclide analysis to derive rates in rapidly denuding terrain, with an investigation of denudation measured using cosmogenic radionuclide analysis in a recently formed mountain range. The assumptions inherent to the techniques of cosmogenic radionuclide analysis have been discussed in relation to the application of the basin-wide approach in areas of rapid denudation. It has been shown that the inclusion of muogenic production becomes necessary in areas where denudation rates exceed $10^{-3}\ \text{mm/a}$. The shielding effect of slope and snow cover will be more pertinent to studies employing cosmogenic radionuclide analysis in areas of high relief, or at high latitudes. In the San Bernardino Mountains, the combined production rate scaling factor for these effects is minimal. Variations in the earth’s geomagnetic field may have a greater significance for production rate scaling. However, the disagreement between results from different studies as to the direction and degree of scaling which must be applied precludes inclusion of such effects until a more comprehensive global palaeogeomagnetic data set exists.
Testing the assumptions of sediment mixing which are required by the basin-wide approach shows that for high relief topography, such as the San Bernardino Mountains, mixing over short length scales is sufficient to provide an integrated sediment sample for analysis. However, at confluences between low order channels, where sediment is being combined from geomorphologically distinct sub-basins, the ratio of mixing may not be representative of the entire upstream basin. Sampling strategies should, therefore, avoid sampling directly downstream of channel junctions and select areas which are geomorphologically homogeneous.

The results reported in this study confirm the feasibility of applying basin-wide cosmogenic analysis in mountainous catchments and show how it is possible to investigate controls of denudation in areas where the influence of climate, crustal motion and lithology can be considered individually. In particular, these controlling factors are considered in regions of the San Bernardino Mountains where landsliding is occurring and where it is not. In landsliding areas, adjacent to the main trace of the San Andreas Fault, any influence of the spatial variations of climate are overwhelmed by the effects of base level lowering though relative crustal uplift. In the non-landsliding areas within the San Bernardino Mountains there is a lack of any climatic or related environmental influence on denudation rates. In the area of the mountains where the low-relief, pre-orogenic form of the landscape has yet to be significantly incised, denudation rates appear to be a function of the changing morphology of slopes over time. Although a relationship exists between the mean slope angles of basins and their denudation rates, this breaks down when mean slope
gradients exceed ~30°. The basins with mean slope angles >30° are those where landsliding is a significant agent of denudation and this suggests that as landscapes reach their threshold angles of repose, denudation rate becomes independent of slope angle. Instead, denudation occurs principally by continuous slope failure in response to base-level drop at the foot of slopes. In small landsliding basins, the mean denudation rates, measured over short and intermediate timescales using sediment yields and cosmogenic radionuclide analysis, approximate the denudation rates recorded at longer temporal and broader spatial scales from thermochronometry. The variance in denudation rates at smaller spatial and temporal scales appears to reflect a greater sensitivity to geomorphic thresholds, rather than reflect forcing by external factors, and it is shown that while the influence of geomorphic thresholds can be discounted in more slowly denuding parts of the mountains they become significant in landsliding areas.

Previous studies have derived a map of denudation rates for the San Bernardino Mountains, averaged over the last 1.25 to 2.5 Ma, from thermochronometry and incision into a marker horizon. The pattern of denudation rates over the last 10²-10⁴ years, derived using cosmogenic radionuclide analysis from small basins, and the long-term rates agree, suggesting a time-transgressive landscape response with distance from the main trace of the San Andreas Fault. This indicates the patterns of denudation in the San Bernardino Mountains have not changed significantly since the inception of their formation. Furthermore, short-term denudation rate measurements over 10¹ years from sediment yield studies in the San Gabriel
Mountains, adjacent to the San Bernardino Mountains, imply an agreement exists between rates of denudation measured over periods of decades, thousands and millions of years on the San Andreas Fault. That denudation rates recorded over a variety of temporal scales appear consistent is not unique but seems to be typical of other mountain ranges, both in southern California and in other regions of the globe. The consistency of denudation rate results over time supports the initial spacing of major drainages as a future control of the topography of the mountains. In ranges where structurally controlled drainage networks may not be able to re-organise, such as appears to be the case in the San Bernardino Mountains, it may be the dimensions of the crustal blocks comprising the range which will define the future form of the mountains. As such, the tempo at which mountains evolve may be as much due to the setting of initial conditions during the early stages of orogenesis rather than the subsequent interactions of tectonic and climatic forces.
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Appendix 1

Sample preparation protocols

A1.1 Introduction:

These protocols describe the sampling methods and sample preparation procedures for cosmogenic radionuclides Beryllium-10 ($^{10}$Be) and Aluminium-26 ($^{26}$Al) used in the Edinburgh University Cosmogenic Laboratories. These protocols have been adapted from those of Ivy-Ochs (1996) and Bierman et al. (2002) and through personal communication with W. Phillips, C. Fogwill, J. Everest, A. Davidson and D. Hughes. Experience has shown that sample preparation should be tailored depending on the lithology of the sample site and the purposes for which cosmogenic samples are being employed. For this reason these protocols, illustrating the processing required to produce targets for accelerator mass spectrometry analysis from alluvial channel sediments (Bierman and Steig 1996) of the San Bernardino Mountains, California, may vary from those of other workers at the University of Edinburgh Cosmogenic Laboratories. In particular these protocols have been developed in order to prepare large (>50 g) quartz samples from bulk sediment containing ~20% quartz.

In the following discussion several points should be assumed. All 2.5 l bottles, 1 l Nalgene bottles, pipette tips, disposable pipettes, centrifuge tubes and caps, Savillex or Teflon beakers and lids, quartz crucibles and their caps are acid cleaned prior to use. All the procedures from stage A1.6 onward are carried out in labs with positive
pressure and High Efficiency Particulate Arrest (HEPA) filtered airflow in which no boro-silicate glass is used. That the labs with HEPA filtered airflow are treated as ‘clean’ labs where lab coats, slippers, gloves, wrist covers and safety goggles are worn at all times. That safety goggles, lab coats, disposable aprons, gloves and heavy acid resistant gloves are worn during stages A1.3 and A1.5.

When referring to purified water the resistivity will be given. All acids used from stage A1.5 onwards are Certified Grade unless stated as Primar Grade.

**A1.2 Sample acquisition:**

In the majority of cases in the San Bernardino Mountains the channels are ephemeral and so alluvial sediment samples collected during dry periods of the year can be sieved in the field to the desired grain size of 125 to 250 µm. This grain size should be large enough to prevent incorporation of wind blown sediment but small enough that it reduces the possibility of non-quartz inclusions within the final, etched, quartz grains. Several studies have found no significant bias of nuclide concentration with grain size (Granger *et al*., 1996; Riebe *et al*., 2000; Clapp *et al*., 2000; 2002; Rienhardt 2003 pers. comm.), while others have reported such a bias (Brown *et al*., 1995; 1998). After finding no significant difference between the nuclide concentrations of one sample for which both grain sizes of 125-250 µm and 250-20000 µm were analysed it is assumed that, at least within the sand size particle range, the grain size selected will have no affect on the denudation rates derived (Chapter 4). Around 1.5 kg of the desired grain size is sieved from the top few tens
of centimetres of alluvial sediment deposited in channel pools or bars and attempts made to localize collection to an area of a few metres. Typically, all the sample material is collected from a single bar or pool. The channel sediment in the small catchments of the San Bernardino Mountains appears sufficiently mixed in channel reaches to prevent significant variation in nuclide concentrations over channel lengths of a few metres (Chapter 4). While some have advocated a more diffuse sampling methodology for larger catchments (e.g. Clapp et al., 2002) for small, steep mountainous catchments where the contributing area may vary significantly over distances of tens of metres up and downstream the above strategy, this is appropriate. In some situations care must to be taken to minimize the possibility that the sediment sampled has originated from bank collapse or other localized mass movement processes.

In cases where channel flow is not ephemeral or when sampling during wetter periods, sieving in the field may not be possible and so bulk samples of around 4 to 8 kg are collected, dried and sieved to the desired grain size in the labs.

**A1.3 Initial etching:**

The aims of the next three stages are to extract pure quartz from the sediment sample and remove any atmospheric, or ‘garden variety’ $^{10}$Be which may be adsorbed to the surface of the quartz grains (Kohl and Nishiizumi 1992). This is achieved by separating and etching non-quartz components and acid cleaning, or etching, the surface of the quartz grains.
The sieved sample is divided into aliquots of ~70 g and placed in several 1 l Nalgene, wide neck, screw cap bottles which are topped up in a fume cupboard with a weak hydrochloric acid (HCl) solution (‘old’ column waste can be used for this purpose) and placed overnight on a shaker table. This helps remove carbonate (CO$_4$) and ferrous (Fe) deposits from the surface of the sample grains, although care has to be taken as carbonates in the sample may react violently with HCl. After this treatment the sample is thoroughly rinsed in tap water, which both removes any organic mass which has not dissolved and is used to float the majority of the mica content, allowing separation and removal of biotite and muscovite.

Without removing any sample the 1 l Nalgene bottles are topped up in a fume cupboard with diluted hydrofluric acid (HF) (1%) and nitric acid (HNO$_3$) (1%) in a fume cupboard and returned to the shaker table for ~1-3 days after which they are rinsed and the process repeated for another ~1-3 days. These two ‘junk’-etches significantly increase quartz concentrations of samples (in the case of the San Bernardino Mountains from ~20% to as much as 75%) so less mass has to undergo secondary etching and the amount of time and space that the sample requires in ultrasonic baths is reduced (see section A1.5). The HF from these treatments destroys the majority of the feldspars present in the sample, which in the case of San Bernardino Mountain granites may be as much as ~70% of the original mineral mass. The HNO$_3$ appears to prevent the formation of a silica gel from the breakdown products, although some workers successfully omit this addition (Stone 2003, pers. comm.)
After the second junk-etch the sample is thoroughly rinsed in water and dried in an oven overnight.

**A1.4 Secondary etching:**

~25 g of sample is weighed into a 2.5 l acid resistant bottle (the 2.5 l container bottles that HF is delivered in are suitable for this purpose). The bottles are filled with 2250 ml of 1% HF and 1% NHO$_3$ in a fume cupboard using a solution of analytical grade HF and HNO$_3$ solution and 15 MΩ H$_2$O. The bottles are left overnight in heated ultrasonic baths set using a timer to alternate half-hourly between on and off for 18 hours (9 hours ‘on’ time per etch) this helps prevent the baths from overheating and the water level from dropping low enough to cause damage to the baths. The next day the samples are rinsed thoroughly, again in a fume cupboard, and the process repeated for a total of three etches. The last etch is performed with a solution using 18 MΩ H$_2$O. It was found that, in samples which had not been separated with heavy liquid, a deep red/purple discolouration of the acid solution used to etch the samples was often observed. This contaminant was not removed by further etching and heavy liquid separation was required.

After the third etch the samples are dried in an oven overnight and the dried quartz bagged and stored.

**A1.5 Density separation and final etch:**

The sample is spread thinly over the bottom of a clean spill tray and a hand magnet is passed over it to extract ferrous material.
Quartz is separated from denser material using heavy liquid. ~100 ml of LST (lithium heteropolytungstate) is placed in a funnel, sealed with a rubber bung. 20-50 g of sample is added, although more LST in a larger funnel would allow greater sample mass to be used. By floating the quartz, any mafic mineral grains present can be sunk and filtered off. Placing a rubber bung in the top part of the neck of the funnel allows segregation of quartz from the mafic content. For some of the initial samples this stage was replaced by magnetic separation in a Franz. However, non-magnetic crystals were not removed by this technique and a similar appearance to quartz and resistance to HF, were not noticed until post quartz dissolution (see section A1.7), whereby physical extraction was required. Hence, the heavy liquid separation technique is preferable at this stage of the sample preparation. By placing it after rather than before the initial etching the amount of material which must be separated is more manageable.

After separation the quartz is thoroughly rinsed and dried in an oven overnight.

The dry, separated quartz is placed in a 1 l Nalgene bottle and etched overnight in an ultrasonic bath using 1%HF (Certified Grade) and 1% HNO₃ (Certified Grade) diluted in 18 MΩ H₂O to remove any residues of LST. After rinsing the sample is dried in an oven.
A1.6 Aluminium assay:

The aluminium assay stage is not essential to the sample preparation but gives a good indication of how pure the sample is and so whether or not it needs further etching before being dissolved.

Once seven quartz samples, of at least 50 g for regions of high denudation, have been collected an aliquot of ~1 g is removed from each, weighed accurately and placed within a 50 ml, acid cleaned Savillex beaker. The amount of quartz required can be approximated using an estimate of the denudation rates and a specific carrier mass.

Several ml of HF (48% Certified Grade) is used to dissolve the quartz aliquot overnight on a hotplate set to 90°C.

Once dried the sample is taken up in 2 ml of 3 M HNO₃ and transferred to a 100 ml measuring flask and diluted with 18 MΩ H₂O to make up a solution of ~100 ml. 1 ml of this solution is further diluted by transferring it to a second measuring flask and filling it to ~100 ml with 18 MΩ H₂O. An aliquot of 30 ml is then taken from the second measuring flask and the aluminium content of this solution measured using Atomic Absorption Spectrometry (AAS) or Inductively Couple Plasma Spectrometry (ICP). The remaining solutions are decanted back into their respective beakers and dried down.
The aluminium content of each sample should be <250 ppm. If this is not the case further etching may bring the value down and one or two etches (depending on the concentration) of the whole sample in a single 2.5 l bottle overnight may resolve this. If the Al concentration is >1000 ppm several more etches may be required, although in some cases the aluminium content of the quartz is so high that <250 ppm concentrations are not possible.

**A1.7 Quartz dissolution:**

The quartz samples are accurately weighed, the weights recorded and the sample placed in 180 ml, acid cleaned, wide mouth, screw cap Savillex beakers which have been labelled with the sample name at least twice on the beaker and once on the lid. To prevent the quartz grains climbing the walls of the often very static Savillex beakers, several drops of 18 MΩ H₂O should be added to weigh down the sample during the next few stages (this must be performed after weighing the dry quartz).

The samples are placed within a fume cupboard and because ⁹Be does not occur naturally in the environment a Beryllium standard solution added. It is at this point that the ¹⁰Be/⁹Be ratio is being ‘set’. There are usually a total of eight Savillex beakers, seven samples and one sample blank. The blank is used to quantify the amount of ¹⁰Be loading incurred during the post-etch sample preparation. The quantity of standard required depends on the expected nuclide concentration of the samples but between 150 and 500 µg of standard (or carrier) has been successfully used in Edinburgh University Cosmogenic Laboratories. The carrier added at Edinburgh University has, up till the time of writing, typically been purchased
(1000±5 ppm Be in 0.5 M HNO₃). However, purchased carrier may contain a background $^{10}\text{Be}/^{9}\text{Be}$ concentration on the order of $10^{-14}$ (Bierman et al. 2002) which can equate to, or be greater than, the radionuclide concentration of rapidly denuding terrain. To resolve this, the amount of carrier added can be reduced, thus reducing the ratio of in situ-produced cosmogenic $^{10}\text{Be}$ to the $^{9}\text{Be}$ added in the carrier and making the blank subtraction less substantial. The same ratio reduction can be achieved by increasing the mass of quartz used, hence increasing the amount of cosmogenic $^{10}\text{Be}$. Reducing the carrier mass is not only likely to give reduced accelerator mass spectrometry beam currents but becomes difficult to handle below ~200 µg, especially if not all of the carrier added is recovered at the end of sample processing. Increasing the mass of quartz used can be time consuming. A prepared using deep mined Beryl can have lower $^{10}\text{Be}/^{9}\text{Be}$ ratios which will help reduce the amount of $^{10}\text{Be}$ loading during sample preparation, making for more accurate measurement of rapidly denuding landscapes. If cosmogenic $^{26}\text{Al}$ is also being measured an aluminium carrier of ~1 ml should also be added.

The Savillex beakers are placed on a hotplate set to 90°C in a fume cupboard and HF (48% Certified Grade) and HNO₃ (70% Certified Grade) are added to the beakers containing the samples in a ratio of 4:1 respectively until the quartz is completely dissolved, for 50 g samples this may take up to 10 days. Care must be taken here as some samples appear to react to the addition of concentrated HF. It is desirable to leave the caps on the samples during this stage with a hole drilled through to allow acid addition and prevent pressure build up. Most of the sample preparation can be achieved with the drilled cap in place but it has to be removed at
certain stages to allow the inside walls of the Savillex beakers to be washed down helping to maximize sample recovery. Also it may be advantageous for larger samples (>40 g) to be stirred every few hours with acid resistant rods.

Note that from this point on the samples contain an appreciable amount of Be which is carcinogenic and once in a dried form poses a serious health risk if inhaled. Any pipette tips, paper towels etc., which come into contact with the samples are double bagged before being disposed of and spills cleaned with copious amounts of water before they have the chance to dry.

A1.8 Fuming:
After all the quartz in the beaker has dissolved a crust of material is left at the base which can range in colour from off white to dark brown. At this point the fume hood where the fuming is going to take place must be washed down as perchloric acid (HClO₄) is used which can form an explosive compound when exposed to organic materials. The samples can be moved to a separate fume cupboard if necessary.

In order to remove fluorides from the samples they are fumed with four additions of 2.5 ml of HClO₄ (70% Certified Grade) (Yokoyama et al. 1999) in a cleaned fume cupboard. This is done using a hotplate set to 220°C. During this stage the beaker caps are removed so that the acid additions can be dripped down their inside walls, washing down any sample precipitated on the wall or which has climbed up the beaker due to static. However, as dried sample can ‘jump’ because of static and
convection currents incurred by heating the sample lids are replaced after acid additions to maximize sample yields and more importantly prevent cross contamination between samples.

After four acid additions and dry downs the beakers are transferred to another fume cupboard and the hood where fuming has taken place must be washed down. Before moving the samples from the cooled hotplate, 2.5 ml of HNO₃ (35%, diluted in 18 MΩ H₂O from 70% Certified Grade) is dripped down the inside walls of the beaker to weigh the sample down during movement, prevent static ‘jump’ of the dried sample material and as a precursor to the Aluminium aliquot stage. Note, any spills of HClO₄ acid must be cleaned using a wet paper towel and rinsed copiously with tap water to prevent the build up of any explosive matter.

**A1.9 Aluminium aliquot:**

In order to quantify the aluminium content of the samples an aliquot is taken, diluted and analysed by AAS or ICP.

In the interests of accuracy each sample is dealt with individually. A 100 ml flask, labelled with the sample name, is tarred on an analytical balance

The sample, taken up in HNO₃ (section A1.8), is transferred to a 100 ml measuring flask in a fume cupboard and topped up to ~100 ml with 18 MΩ H₂O. Several 5 ml additions of the H₂O being used to dilute the sample solution are used to rinse the inside of the beaker to make sure all the sample is transferred to the measuring flask.
After several water rinses any solid material, which should not be transferred from
the beaker to the flask, must be removed and stored in a labelled vial for later
analysis. The flask is inverted several times to allow mixing, reweighed and the
weights recorded.

Using a second labelled and tarred 100 ml measuring flask, 2.5 ml of the solution in
the first flask is transferred to the second, reweighed and the weight recorded.

The second flask is topped up to ~100 ml using 18 MΩ H₂O, inverted several times,
reweighed and the weight recorded.

~30 ml of the solution from the second flask is decanted into two labelled bottles,
one for AAS or ICP and one as an archive.

The remaining contents of both flasks is decanted back into respective Savillex
beakers and dried on a hotplate at 90°C.

**A1.10 Anion columns:**

Any Fe in the quartz can be dealt with by using anion exchange columns whereby
resin in the column is treated with strong HCl coating the surface of the resin with
hydrogen (H) cations, which adsorb anions. Fe in the sample solution forms a
chloride anion in HCl and when dripped through the columns containing resin is
held on the resin while the Be and Al cations pass through freely. Introducing a
weak HCl solution desorbs the Fe anions which can be eluted into a waste container.
Batches of eight samples are typical. 20 ml columns containing Anion Exchange Resin (Bio-Rad AG1-X8, 100-200 mesh, chloride form) which have been preconditioned with 18 MΩ H₂O by the last user are set up in a rack at a height appropriate to catch any elutant.

Stock solutions of 9 M HCl and 0.5 M HCl are made up using 37% Primar Grade HCl and 18 MΩ H₂O. This stock solution is used to fill eight bottles (marked A) with 60 ml 9 M HCl; one bottle (marked B) with 40 ml 9 M HCl; eight bottles (marked C) with 40 ml 9 M HCl; eight bottles (marked D) with 80 ml 0.5 M HCl. Eight bottles (marked E) should also be filled with 60 ml 18 MΩ H₂O. One of each of the bottles of A, C, D and E will be eluted through each of the eight columns.

5 ml of acid from bottle B is added to each sample beaker to take up the sample in solution, this will likely be an orange or yellow colour due to the Fe and other anions in the sample.

The contents of bottle A are added to each of the respective columns with the elutant captured in a labelled ‘waste’ bottle.

The sample solution is added to each of the respective columns taking care to record which sample is in which column.
Working as quickly as is safe, several ml from bottle C are used to rinse the Savillex beaker and add the rinsed solution to the respective column. This should be done one or two times but quickly enough that the Be and Al are still near the top of the column.

The ‘waste’ bottle is replaced by the Savillex beaker and the remainder of bottle C is added to the respective columns.

Once bottle C has been added the sample should have been pushed through the column minus any anions which are still be adsorbed to the resin surface, the eluted sample should be colourless. The beaker can be removed and placed on a hotplate in a fume cupboard to dry at about 90°C. The ‘waste’ bottle can be used to collect the elutant as bottles D and E are added to the columns and a yellowish discolouration may be noticed here. Note the ‘waste’ bottles should be carefully labelled as they may contain the sample if mistakes have been made and recovery should be possible. Before the last of bottle E has gone through the columns they should be capped to prevent the resin from drying out.

**A1.11 Titanium precipitation:**

Titanium (Ti) reduces beam currents during analysis by accelerator mass spectrometry but can be removed from the sample by precipitating it out of the sample solution at a pH of ~4.
Two 10 ml additions of 1.2 M HCl (Primar Grade), are used to transfer the samples to 50 ml centrifuge tubes labelled with the sample name and “tube 1”.

30% ammonia (NH$_4$OH) is used to bring the pH of the solutions to between 3.8 and 4.2 (use a 1.75 ml addition initially and then add drop by drop), they are then centrifuged at 3000 rpm for 10 minutes.

Any precipitate in the base of the centrifuge tube is Ti. Note, that some have reported loss of Be due to incorporation in the Ti flocculate during this procedure (Child et al. 2002). However, Ti may not always be present and so evidence of a precipitate is not mandatory.

The supernatant, which contains the Al and Be, is decanted into a second 15 ml centrifuge tube labelled with the sample name and “tube 2”.

By bringing the pH of this solution up to 9, a flocculate should be observable, although for samples with low Al concentration and small Be carrier it can be difficult to observe this immediately.

Letting the tubes sit overnight at this stage is reported to help reduce $^{10}$B levels (Ivy-Ochs 1996).
Samples in tube 2 are spun at 3000 rpm for 10 minutes and the supernatant poured off into tube 1. There must be a noticeable precipitate at this stage or the sample has been lost (the anion waste bottle may contain the sample).

3 ml of 18 MΩ H₂O is added to each centrifuge tubes, shaken and centrifuged at 3000 rpm for 10 minutes, the supernatant is poured off into tube 1 and archived. Again the sample must be visible in centrifuge tube 2. The water wash helps clean impurities from the sample.

3 ml of 3M HCl (Primar Grade) is added to each tube 2, shaken and once the samples are fully dissolved they are poured back into respective beakers.

These tubes are then rinsed with 2 additions of 1 ml 3 M HCl adding it back into the respective beakers.

The samples are dried at 90°C on a hotplate in a fume cupboard.

**A1.12 Cation columns:**

The resin in the cation columns desorbs Be and Al at different acid strengths allowing their separation and hence requires that an extra beaker, or bottle, is available for each pair of Al and Be expected to be measured by accelerator mass spectrometry.
Columns are usually done in batches of eight samples. 20 ml columns containing Cation Exchange Resin (Bio-Rad AG50W-X8, 100-200 mesh, hydrogen form) which have been pre-conditioned with 18 MΩ H₂O by the last user are set up in a rack at a height appropriate to catch any elutant.

Stock solutions of 9 M HCl, 4.5 M HCl and 0.5 M HCl are made up using 37% Primar Grade HCl and 18 MΩ H₂O. This stock solution is used to fill eight bottles (marked A) with 60 ml 9 M HCl; eight bottles (marked B) with 60 ml 4.5 M HCl; eight bottles (marked C) with 60 ml 1 M HCl; eight bottles (marked D) with 60 ml 1 M HCl; eight bottles (marked E) with 160 ml 1 M HCl; eight bottles (marked F) with 80 ml 1 M HCl; eight bottles (marked G) with 80 ml 4.5 M HCl. Eight bottles (marked H) should also be filled with 60 ml 18 MΩ H₂O. One bottle (marked J) should be filled with 40 ml 1 M HCl. One of each of the bottles of A, B, C, D, E, F, G and H will be eluted through each of the eight columns.

5 ml from bottle J is added to each of the sample beakers to take up the sample into solution.

The contents of bottles A, B and C are added to each of the respective columns with the elutant captured in a labelled ‘waste’ bottle.
The contents of the sample beakers are added to each column.

The contents of bottle D can be used to rinse the inside of the sample beakers and this solution added to the columns. This is done several times to maximize the yields and help achieve a clean separation of Be and Al. After bottle D has been added to the columns the elutant will caught in the Be sample (Savillex or Teflon) beaker.

The contents of bottle E are added to the columns. After bottle E has been added to the columns the sample beakers containing the Be are moved to a fume cupboard and dried down at 90°C and the elutant captured in the ‘waste’ bottle.

The contents of bottle F are added to the columns. After bottle F has been added to the columns the elutant is caught in the Al sample (Savillex or Teflon) beaker.

The contents of bottle G are added to the columns. After bottle G has been added to the columns the Al sample beaker is moved to a fume cupboard and dried down at 90°C and the elutant captured in the ‘waste’ bottle.

The contents of bottle H are added to the columns, to condition them for the next user. Note, the ‘waste’ bottles should be carefully labelled as they may contain samples if mistakes have been made and it should be possible to recover them.
Before the last of bottle H has gone through the columns they are capped to prevent the resin from drying out.

**A1.13 Final fume:**

A final fume in HClO$_4$ helps reduce $^{10}$Be levels, hence improving accelerator beam currents (Bierman 2002).

The Be sample beakers have 1 ml of HClO$_4$ added to each and are placed on a hotplate set to 250°C in a fume cupboard which has been thoroughly washed down to prevent formation of explosive compounds.

After the samples have dried down completely and cooled, 2 ml of 3 M HCl is dripped down the insides of each beaker to maximize sample yields and as a precursor to the hydroxide precipitation stage. The fume cupboard should be washed down thoroughly.

**A1.14 Hydroxide precipitation:**

Aluminium and Beryllium hydroxide gels are precipitated out of solution in order to prepare targets to press.
In a fume cupboard, the sample solutions are transferred to labelled 15 ml centrifuge tubes. Two, 2 ml 18 MΩ H₂O additions are dripped down the insides of the beaker to rinse all of the sample from the beaker into the centrifuge tube.

30% NH₄OH is used to bring the pH of sample solutions up to ~9, where they are shaken and centrifuged at 3000 rpm for 10 minutes. A precipitation must be observed in the tubes.

The supernatant is decanted into the respective sample beaker and the samples dissolved in 2 ml 18 MΩ H₂O to which a drop of HCl (37% Primar Grade) has been added, they are then shaken. Adding drops of 3% NH₄OH, the samples should be precipitated, shaken and then stored overnight to allow all the ¹⁰Be to come out of solution.

The samples are then centrifuged at 3000 rpm for 10 minutes. This double precipitation helps rinse the sample of contaminants.

The supernatant is poured off into the respective beakers, note the supernatant is a gel and so should stay at the end of the tube allowing the supernatant to be easily tipped out of the tube. 5 ml of 18 MΩ H₂O is added to the samples which are shaken and centrifuged at 3000 rpm for 10 minutes. The supernatant is poured off and another 5 ml 18 MΩ H₂O is added to the samples which are shaken and
centrifuged at 3000 rpm for 10 minutes. This double rinse helps remove impurities from the samples and allows them to form solid ‘pellets’ of hydroxide after they have dried instead of a thin scum at the base of the centrifuge tubes. These pellets are easier to transfer to quartz crucibles, easier to press into targets and run less risk of sample loss during the packing and pressing stages.

The centrifuge tubes are carefully cut (at around the 5 ml mark) making sure they still labelled and no plastic tube falls into the sample gel. This provides a more rapid dry-down and makes the transfer of the sample to the quartz crucibles easier.

The samples are placed on a block heater set to 70°C in a fume cupboard and heated until dry.

**A1.15 Firing:**

In this last stage the dried hydroxides are oxidized and packed ready for sending to be pressed and analysed. Some accelerator facilities may prefer the sample to be mixed with copper (Cu) or Niobium (Nb) powder, in the case of $^{10}$Be and sliver (Au) powder, in the case of $^{26}$Al, at this stage. Most will be willing to perform this step, however.

Eight quartz crucibles are taken directly from their acid cleaning treatment and stored under 18 MΩ H$_2$O until needed.
A quartz crucible is selected, fired over a gas flame for several seconds till completely dry and sterile and then placed in a labelled holder to cool, once cooled it is capped and weighed on an analytical balance and the weight noted.

Working over a containment tray the pellet of Be[OH]$_2$ or Al[OH]$_3$ is tipped into the quartz crucible and fired over a gas flame until the pellet glows cherry red for about 30 seconds. It is returned to the holder, left to cool, capped and reweighed to obtain the weight of the BeO or Al$_2$O$_3$.

The crucibles are then placed in labelled glass vials to be shipped to the laboratory where they are to be pressed into targets.
Appendix 2

Basin averaged cosmogenic radionuclide production rates

A2.1 Introduction:

The derivation of realistic radionuclide production rates is fundamental to the use of the cosmogenic technique in earth science applications. For alluvial samples, amalgamating sediment denuded from across whole basins, there must be an appropriate averaging of production rates. Here, a method for calculating basin averaged cosmogenic $^{10}$Be production rates from digital elevation models using ArcView 3.x computer software is presented. The same methodology can also be applied to cosmogenic $^{26}$Al if the appropriate variables are known. The scaling factors of Lal (1991) are used to calculate the spallation component of production and Stone (2000) Granger and Smith (2000) and Granger et al. (2001) are used to derive the fast and slow muogenic components. The scaling method of Dunai (2000) is also possible using a variant of this method. Note that the method outlined below scales production rates only for latitude and altitude and that the affect of geomagnetic field variation and sample shielding issues are not accounted for (chapter 3.4). While the method is presented in the form of a 'recipe', some knowledge of ArcView and the way in which the software file structure works is assumed, as is the knowledge needed to acquire digital elevation models. When referring to ArcView menus and functions italics will be used.
A2.2 Setting up ArcView:

The ability to manipulate digital elevation models in ArcView requires the Spatial Analyst extension be loaded and turned on. Here we will use several ArcView scripts, chunks of Avenue code that add extra functionality to the software. The first is a powerful hydrological extension\(^{13}\). This will add several buttons and menu tabs to the ArcView project and will allow definition or 'cookie-cutting' of the basin areas of interest. Other scripts are available to achieve this function and could be incorporated just as easily into the procedure outlined here, although this script appears to be very stable while others which have been tried have failed. Download the file from the authors web site and follow the instructions to save the project with the Hydrology functionality added\(^{14}\).

Next go to ESRI scriptsearch website\(^{15}\). This facility allows the web to be searched for Avenue code written by other ArcView users, with which it is possible to customise a particular project. The Grid Properties script\(^{16}\) should be downloaded which allows basic statistical characteristics of digital elevation models to be calculated and, if United States Geological Survey National Elevation Dataset digital elevation model data is being used, the Grid Projector script\(^{17}\) should be downloaded. Searches for these scripts should be limited to Avenue code and ArcView GIS software. It is advantageous to create a separate folder in which scripts and downloaded zip files can be saved. To load these scripts into the project

\(^{13}\) Written by Bernie Engel.
\(^{14}\) See http://pasture.ecn.purdue.edu/~sengelb/abe526/wshddelin/wshddelin.html
\(^{15}\) See http://arcscripts.esri.com/
\(^{16}\) Written by William Huber.
double click *scripts* from the *project* window, select *load text file* from the *script* tab menu and *compile*. The script should be renamed to something appropriate using the *script* tab and *properties*, then the script window closed. By clicking on *customise* in the *project* tab a button for the script can be created. This will appear in the menu or toolbar specified. To do this change the `type' to view and the `category' to button, click `new' to add a blank button and double click `click' in the list at the bottom to select the script to add as a button. Double click the `icon' command in the list to give the button a design. The hydrological extension script will add menus and buttons without user intervention.

Once the scripts have been created it is appropriate to create two directories for this project. One temporary, to which the working directory can be set, and one where permanent files can be stored once it has been decided they will be kept. *Save as* the project in an appropriate file. Digital elevation model data can now be added to the project. Create a new *view* into which the digital elevation model grid should be added using *add theme*. If United States Geological Survey National Elevation Dataset data is being used create a separate view and convert the digital elevation model into metres using the *Grid Projector* button. The need for this stems from the fact that United States Geological Survey digital elevation model data is projected in decimal degrees in the x and y but in metres in the z plane which prevents hillshading. Reprojecting all the grid axes into metres solves this but note that most of the calculations will be performed on the decimal degree grid and only need to be reprojected into metres for display purposes.

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*Written by Kenneth McVay.*
A2.3 Calculating Production Rates:

A2.3.1 Defining the Catchment:

Here, the basins of interest from the digital elevation model are defined. For this the hydrological extension script is used. Make the (decimal degree) digital elevation model active and in the Hydrology menu select fill sinks selecting the digital elevation model grid as input when prompted. This may take some time to run depending on the size of the digital elevation model. When done select flow direction from the same tab and then do the same with flow accumulation. Each time prompting will be given for the input grid and a name for the output grid. The next step is to define the streams based on the flow accumulation grid from the option in the Hydro tab menu, giving the stream grid a name and choosing the number of cells in the digital elevation model which define the upstream area. This value will depend on the digital elevation model you are using and the stream density required, hence a little experimentation is advisable. Making the digital elevation model grid active, press the watershed delineation button. Position the cursor at the sample point, i.e. the head of the basin of interest, placing it over the stream defined by the digital elevation model if possible. When the cursor is positioned clicking will define the basin and it will appear as a grid. This is the `mask' for that particular basin, rename it something appropriate. If the basin has not been correctly defined it may be because of artifacts or errors in the digital elevation model data. Try again but a little further upstream, unless the basin being defined is very small this will likely produce negligible error in the final production rate.
In this discussion the procedure for production rate calculation will be limited to one basin. If there are many basins in a digital elevation model to be defined, use can be made of mosaic grid feature in the theme tab menu to produce a mask of several basins allowing calculations to be performed on all basins at once. This speeds up production rate calculation. However, the single basin masks such as the one just defined should not be deleted as they will be required at the end to separate each basin from the mosaic. If nested basins (i.e. a larger basin split into sub-basins) are of interest each sub-basin mask must be defined but production rate calculations must only be performed once on the whole basin and then split the results for the sub-basins using the masks at a later stage.

A2.3.2 Calculating Latitude:
The scaling factors of Lal (1991), table 2, require latitude as does the muogenic scaling factor given by Stone (2000). For small basins the variation in cosmogenic production rates across the basin may be negligible and an estimate will be sufficient in most cases but for larger areas, or for completeness, ArcView can be used to quickly produce a grid of basin latitudes. Working in the decimal degree view, with the digital elevation model grid active, select properties from the analysis tab and where it asks 'set mask' select the cookie-cut basin grid defined in the last section. This means that ArcView will focus only on the area within this mask (i.e. the basin) and ignore the rest of the digital elevation model. Also, the cell size and grid size must be set to the same as the digital elevation model grid. The grid needs to be integer rather than floating point so select map calculator from the analysis menu.
and double click the digital elevation model grid which appears in the list on the left.
Place the cursor at the end and type .int producing:

((digital elevation model grid)).int and click evaluate. Add the theme to the view
when prompted. Make the integer digital elevation model grid just created active
and in the theme tab select convert to shapefile. The result is a shapefile of cells,
some of which are merged as they had the same elevation but this is unlikely to be
common except in very shallow basins. Make the shapefile just created active and
convert to grid from the theme tab. Prompting will ask which field to use for values
in the grid, select the y coord field and add the grid to the view. There will now be a
grid of latitude values to use in scaling calculations.

A2.3.3 Calculating Atmospheric Pressure:

The scaling functions of Granger and Smith (2000) and Stone (2000) use
atmospheric pressure derived from elevation. This can be calculated by applying the
standard pressure model (Stone 2000), using the digital elevation model elevation
values, in ArcViews map calculator function. When applying this formula in map
calculator it is easier to split it into smaller parts, produce the grids from those parts
and then use those grids in the final calculation. Note, log (Log) in ArcView map
calculator is natural log (ln). The result should be a grid of atmospheric pressure.

A2.3.4 Deriving Spallation Induced Production:

To obtain the spallation component of production rates Lal’s (1991) table 2 is used.
This specifies an a, b, c and d coefficient which, when fitted to a third order
polynomial along with elevation, gives the scaling value for the sea-level high
latitude production rate (for $^{10}$Be assumed here to be 5.1, (Stone, 2000)). However, these coefficients vary with latitude and so if this is considered to be relevant (see above) a grid for each of the four coefficients can be produced which can be entered into the scaling function polynomial in map calculator. To do this, a table of Lal's values for the latitudinal range of interest should be produced using a spreadsheet application and then imported into ArcView. ArcView is unable to operate on tables with decimal places. However, as the latitudinal values are only used to map coefficient values to the appropriate latitude, the decimal point in the table can be moved as long as we do the same in the latitude grid in ArcView. Therefore, multiply latitude in the table of Lal's values produced in the spreadsheet by, for example, 1000. This table should have five columns with headers, the first being latitude (multiplied by 1000, or with no decimal places) then a, b, c and d. As the latitudinal range of the basin is unlikely to be large, interpolation of the coefficient values between the ones Lal gives will be required. The style of interpolation is up to the user though here, as yet, only linear interpolation has been employed. Save the table as a text file, comma or tab delimited. In the project window add table from the project tab should be selected the table just saved located using browse, this will import it. In the view window map calculator is used to produce a latitude grid which is then multiplied by the same value as the latitude values in the imported table (e.g. 1000), adding .int at the end of the calculation to remove any trailing decimal points. Next this grid is converted into a shapefile as above and the open theme table button clicked. Open the table just imported. These two tables must be joined so that the latitude values in the shapefile table (termed the destination) will correspond to the latitude values in the table of Lal's coefficients (termed the
source). To do this select the source table and click on the latitude header. Next, go to the destination table and click on its latitude header, termed 'gridcode'. Click on the join button and the tables will appear to be one. Convert the latitude shapefile into a grid using the convert to grid function in the view window. Prompting for which field to use ‘as values’ will appear, choose 'a' and a grid of Lal's ‘a’ coefficient values can be added to the view, repeating this process for the b, c and d coefficients produces a grid of each.

Use the map calculator function to produce a grid of elevation in km (e.g. divide the digital elevation model grid by 1000 if elevation is in metres). Then, using map calculator the polynomial given in Lal (1991) is entered using the a, b, c and d grids, the elevation grid in km and noting that to cube something in map calculator multiply it by itself three times. The result should be divided by 563.4 which collapses Lal's table 2 so that nuclear disintegration rates are converted to production rates and the result of this is multiplied by 5.1 giving a grid of spallation production rates for $^{10}$Be.

The mean production rate for the basin can be calculated by making the production rate grid active and using the Grid Properties script. If the above procedure has been followed this comprises simply of clicking the Grid Properties button. The mean should be one of the statistics displayed, this is the whole basin production rate of $^{10}$Be by spallation corrected for altitude and latitude.
A2.3.5 Muon Production:

Here the values given in Granger et al. (2001) are used to calculate muogenic production. Two separate production rates for slow muons (representing two different cut-off energies) are required and a production rate for fast muons is also needed to fit the denudation rate equation (section 3.4). Muon production scales with latitude according to Stone (2000, table 1). The importance of latitude scaling for muogenic production is likely to be even less than that of spallation but again, for completeness, a grid of muogenic latitude scaling can be produced using the same method used for Lal's coefficients, above. This means importing a table of latitude and muogenic scaling coefficients from spreadsheet software, joining it to the latitude shapefile table and converting the shapefile containing the new data column to a grid displaying the muogenic scaling values. The correction for altitude, more precisely atmospheric pressure, is done in map calculator using the formula given in Stone (2000) which is: \( e^{1013-P/242} \) (see section 3.4.1.2) where \( P \) is the pressure grid. Multiplying the pressure correction grid by the latitude scaling grid gives a muogenic production rate scaling grid for the basin, the mean of which is derived using the Grid Properties button. This value is then used to scale the \( A_1 \), \( A_2 \) and \( B \) values (Granger et al. 2001) in the denudation rate equation (section 3.3.4). To gain a visual appreciation of the production rate variation across the basin the grids can be draped over hillshades, note if using United States Geological Survey data, conversion of the production rate grid from decimal degrees to metres using the Grid Projector script as noted above will be necessary.
## Appendix 3a

**10Be Production rate scaling values**

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<tr>
<th>Sample ID</th>
<th>Basin averaged elevation (m)</th>
<th>Atmospheric pressure (mbar)</th>
<th>Average basin latitude (dec.deg.)</th>
<th>Basin averaged spallation 10Be production rate (at/g/a)</th>
<th>Basin averaged muon production rate scaling factor</th>
<th>Slope of basin long-axis (°)</th>
<th>Slope Scaled Spallation Production Rate (at/g/a)</th>
<th>Muogenic production rate scaling factor</th>
<th>Snow depth (cm)</th>
<th>6 month spallation scaling for snow</th>
<th>Scaled 10Be spallation production rate (at/g/a)</th>
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Basin-wide $^{10}\text{Be}$ production rates are scaled for latitude and elevation according to table 2 of Lal (1991) (see section 3.4.1.2). The rates are then corrected for shielding by slope-angle (see section 3.4.2.1) and estimated snow cover (see section 3.4.2.2). The high-latitude sea-level $^{10}\text{Be}$ production rate is assumed to be 5.1 at/g/a (Stone 2000). A 10% error is applied to production rate after scaling corrections. Sample MHC-11mean is the average of MHC-11g and MHC-11n. Sample MHC-2mean is the average of MHC-2a, MHC-2b and MHC-2c. Samples MHC-13i and MHC-8i are the denudation rates derived for the basin areas between sample sites MHC-13 and MHC-8 and between sample sites MHC-8 and MHC-2 respectively.
## Appendix 3b
### Laboratory derived values

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<th>Carrier mass (μg)</th>
<th>Carrier mass error</th>
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<th>Number of atoms (at/g)</th>
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For the denudation rate calculation: bedrock density is 2.6±4% g/cm³; attenuation length is 145±3% g/cm² and $^{10}$Be half-life is 1.51 Ma. Values and errors are taken from Gosse and Phillips (2001). $Y_1=5.6\times10^{-4}$; $A_1=170.6$; $A_2=36.75$; $L_1=738.6±3%$; $L_2=2688±3%$; $L_3=4360±3%$; B=0.026 from Granger et al. (2001) with a 3% error added to $L_1$, $L_2$ and $L_3$ in keeping with the error added to the attenuation length of spallation. High-latitude, sea-level spallation production rate is assumed to be 5.1 at/g/a (Stone 2000). Production rates are scaled for altitude, latitude and shielding as discussed in Chapter 3 and given in Appendix 3a. Errors on denudation rate are the propagated 1 standard deviation error of the values given assuming a 10% error on production rates. Sample MHC-11mean is the average of MHC-11g and MHC-11n. Sample MHC-2mean is the average of MHC-2a, MHC-2b and MHC-2c. Samples MHC-13i and MHC-8i are the denudation rates derived for the basin areas between sample sites MHC-13 and MHC-8 and between sample sites MHC-8 and MHC-2 respectively.
## Appendix 3c
### Basin characteristics and sample averaging times

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Basin slopes and areas are derived from 10 m digital elevation model data. Sediment volumes produced are the product of denudation rate and area. The averaging times are calculated using the mean of the averaging times of spallation, slow and fast muons weighted by the relative contribution of each to the total production rate. Averaging times should not be considered absolute periods of time but they give an indication of the timespan of measurement (section 3.5). Sample MHC-11mean is the average of MHC-11g and MHC-11n. Sample MHC-2 mean is the average of MHC-2a, MHC-2b and MHC-2c. As such neither incorporate an averaging time. Samples MHC-13i and MHC-8i are the denudation rates derived for the basin areas between sample sites MHC-13 and MHC-8 and between sample sites MHC-8 and MHC-2 respectively. As such neither incorporate an averaging time.
Further References Cited

