

# Long-term rates of denudation in the Dry Valleys, Transantarctic Mountains, southern Victoria Land, Antarctica based on in-situ-produced cosmogenic $^{21}\text{Ne}$

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Received 14 April 1997; revised 2 March 1998; accepted 15 May 1998

## Abstract

Concentrations of cosmogenic  $^{21}\text{Ne}$  ( $^{21}\text{Ne}_c$ ) measured in quartz have been used to estimate long-term rates of denudation for contrasting landscape components in the Dry Valleys area of the Transantarctic Mountains, southern Victoria Land, Antarctica. Samples of Beacon Supergroup sandstones and granitic basement were collected from two contrasting landscape elements—low-relief, high-elevation surfaces and rectilinear slopes—to assess variations in rates of denudation with topographic position. The sample sites for rectilinear slopes were selected because of proximity to  $^{40}\text{Ar}/^{39}\text{Ar}$ -dated lavas and ash-avalanche deposits. All  $^{21}\text{Ne}/^{20}\text{Ne}$  ratios are significantly greater than the atmospheric value, and concentrations of  $^{21}\text{Ne}_c$  were calculated from the measured  $^{21}\text{Ne}$  values assuming an atmospheric composition for the trapped component. Apparent exposure ages calculated from the concentrations of  $^{21}\text{Ne}_c$ , assuming no denudation since exposure, range from 3.78–4.66 Ma for samples from the high-elevation plateau surface, and 0.61–2.48 Ma for samples from the rectilinear slopes. Exposure ages for  $^3\text{He}_c$  were 2 to 42 times lower than those derived from the abundances of  $^{21}\text{Ne}_c$  because of preferential diffusive loss of  $^3\text{He}_c$  from the quartz lattice; concentrations of  $^3\text{He}_c$  were, therefore, not used in the calculation of rates of denudation. We interpret the abundances of  $^{21}\text{Ne}_c$  as reflecting variations in the rates of denudation rather than exposure age in view of independent evidence for prolonged exposure (> 15 Ma) of bedrock surfaces at the sample sites. Calculated maximum rates of denudation range from 0.26–1.02 m  $\text{Ma}^{-1}$  for the rectilinear slopes, down to only 0.133–0.164 m  $\text{Ma}^{-1}$  for the high-elevation surface sites. These rates are comparable to other estimates of denudation rates for the Dry Valleys derived from analyses of cosmogenic isotopes, but are around two orders of magnitude lower than the long-term mean rate over the past ~ 50 Ma estimated from fission-track thermochronology. Combined with the preservation of volcanic deposits dating back to the mid-Miocene, these  $^{21}\text{Ne}_c$  data demonstrate that only minimal modification of the landscape has occurred in the Dry Valleys over at least the past ~ 15 Ma. This significant conclusion supports the view that

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the East Antarctic Ice Sheet has been essentially stable over this period rather than experiencing major fluctuations as late as the Pliocene, as has previously been suggested. © 1999 Elsevier Science B.V. All rights reserved.

*Keywords:* denudation rates; cosmogenic isotopes; landscape evolution; paleoclimatology; Antarctica

## 1. Introduction

We report quantitative estimates of rates of denudation derived from measurements of in-situ-produced cosmogenic  $^{21}\text{Ne}$  for contrasting landscape elements in the Dry Valleys region of the Transantarctic Mountains in southern Victoria Land, Antarctica. A knowledge of rates of landscape modification in this area is important for several reasons. First, the current debate about the stability of the East Antarctic Ice Sheet has involved contrasting interpretations of the antiquity of the landscape of the Transantarctic Mountains. The model of an unstable ice sheet (Webb et al., 1984; Webb and Harwood, 1991; Barrett et al., 1992) requires much warmer conditions, with temperatures 20–25°C higher than those of the present, as recently as 3 Ma BP. Such a climatic environment implies significant rates of glacial and periglacial geomorphic processes and landscape change. By contrast, those advocating long-term stability of the ice sheet for the past 15 Ma or more have viewed the present landscape as being essentially relict with minimal change under a hyper-arid polar climate similar to that of the present (Clapperton and Sugden, 1990; Denton et al., 1993; Marchant et al., 1993a; Sugden et al., 1995a). Secondly, irrespective of these conflicting interpretations of the late Cenozoic climatic history of Antarctica, the landscape of the Dry Valleys has experienced stable climatic conditions for, at the very least, the past 2–3 Ma; this is in contrast to the glacial–interglacial-driven oscillations in temperature and precipitation that have affected virtually every other morphoclimatic zone. Thirdly, during this period of climatic stability, much of the landscape has developed under an extreme hyper-arid, frigid regime in the virtual absence of running water, a situation encountered nowhere else on Earth. Finally, various lines of evidence, including the preservation of volcanic deposits of known age and already published data from in situ-produced cosmogenic isotopes, suggest that rates of denudation in this area are the lowest for any

terrestrial environment, and, therefore, represent a benchmark against which rates of denudation in areas of greater geomorphic activity elsewhere can be compared.

Over the past decade, measurements of in-situ-produced cosmogenic nuclides in bedrock exposures and surface deposits have begun to provide chronological information on Quaternary, and in some cases, Pliocene events, and valuable site-specific and basin-scale estimates of long-term rates of denudation (Nishiizumi et al., 1991; Bierman and Turner, 1995). The majority of applications of analyses of cosmogenic nuclides in previous studies have involved the age of surface exposure, the calculation of which is based on the assumption of zero denudation since an exposure event. This assumption can be constrained in some cases by independent evidence, or can be reasonably assumed for certain very young surfaces, such as those exposed by the most recent episode of deglaciation in the mid-latitudes. In most situations, however, geomorphological evidence exists for progressive denudation, albeit at an extremely low rate in some cases. Therefore, it is generally more appropriate to interpret cosmogenic isotope data in terms of rates of denudation rather than ‘exposure ages’. Here we use measurements of in-situ-produced cosmogenic  $^{21}\text{Ne}$  in quartz to provide estimates of rates of denudation in the Dry Valleys region (Fig. 1). More specifically, we attempt to characterize differences in rates of denudation between high-elevation surfaces and rectilinear valley-side slopes, the two most significant landscape components in this geomorphologically unique environment.

$^{21}\text{Ne}$  has particular advantages as a cosmogenic isotope for use in environments with potentially extremely slow rates of denudation because, as a stable nuclide, it does not have an upper limit to its accumulation as a result of losses through radioactive decay. Even in the case of  $^{10}\text{Be}$ , which has the longest half-life ( $\sim 1.5$  Ma) of presently used cosmogenic radionuclides, the maximum age of expo-

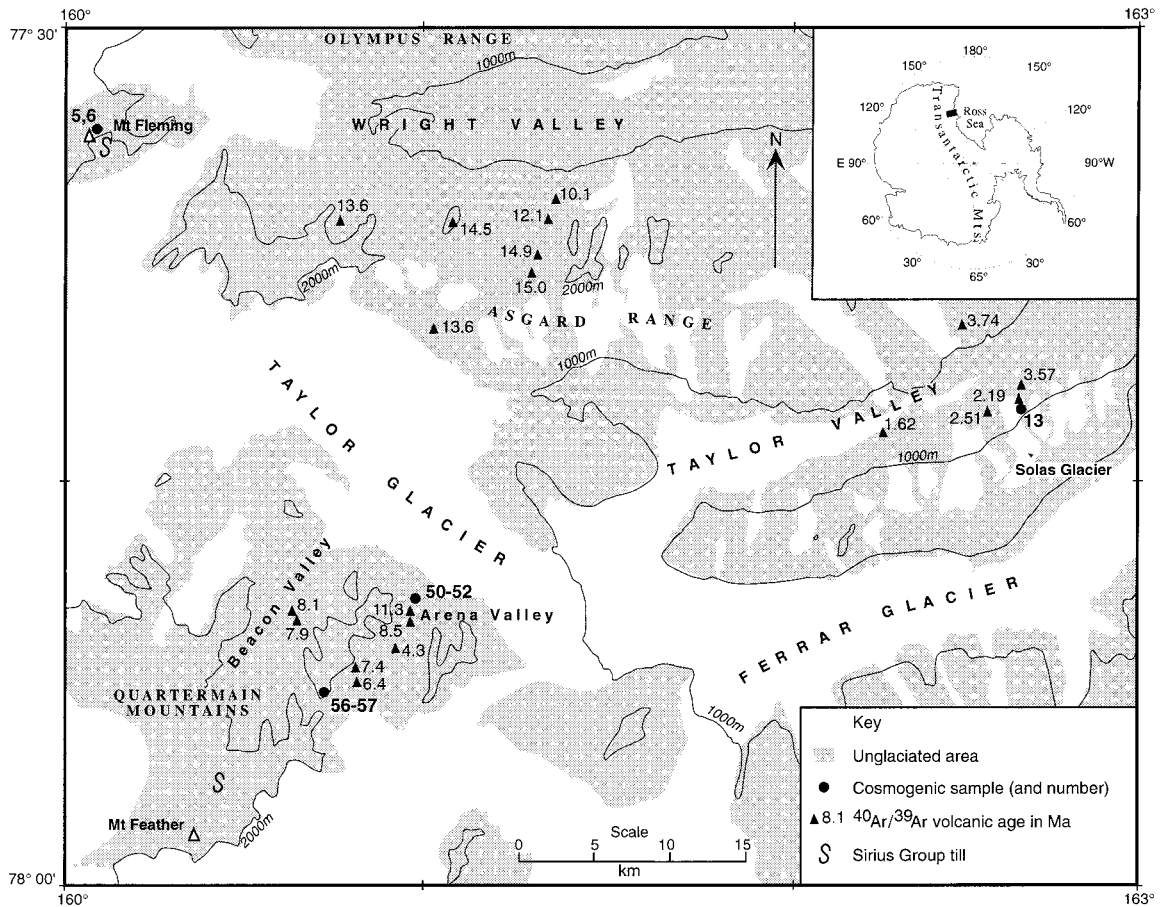


Fig. 1. Map of the Dry Valleys area of the Transantarctic Mountains showing sites sampled for the analysis of cosmogenic  $^{21}\text{Ne}$ . Locations and  $^{40}\text{Ar}/^{39}\text{Ar}$  ages for volcanic deposits are also shown.

sure attainable, assuming zero denudation, is  $\sim 5\text{--}6$  Ma. Whereas the diffusion rate of cosmogenic  $^3\text{He}$  in quartz limits its ability to determine rates of denudation (Trull et al., 1991), diffusive loss of cosmogenic  $^{21}\text{Ne}$  from quartz is minimal and  $^{21}\text{Ne}$  can, therefore, be used to determine denudational histories on timescales of  $10^6\text{--}10^7$  a (Graf et al., 1991).

## 2. Physical setting

### 2.1. Morphotectonics of the Transantarctic Mountains

The Transantarctic Mountains extend for more than 3000 km in a broad arc from northern Victoria

Land on the Pacific coast to the Theron Mountains which terminate near the South Atlantic Ocean. Rather than being a mountain range *sensu stricto*, they consist of a large amplitude ( $\sim 2000\text{--}4000$  m), short wavelength (50–200 km) upwarp forming the rim of an extensive plateau that rises from the interior of East Antarctica (Kerr et al., *in press*). The outer flank of this upwarp is marked by a significant topographic discontinuity comprising either a major escarpment, or a stepped series of minor scarps. This escarpment landscape is most dramatically developed where it fringes the structural basins forming the Ross Embayment.

The main elements of the present structural setting of the Transantarctic Mountains in the Dry Valleys area can apparently be traced to asymmetric

rifting in the Eocene, with extension in the lower plate represented by the Ross Embayment, and block tilting, or flexure, of the upper-plate margin producing a gentle inland dip in Devonian–Triassic age sedimentary units (Fitzgerald et al., 1986; Fitzgerald, 1992). The amount of surface uplift in the Dry Valleys area associated with this rifting event is uncertain; however, results from apatite fission-track thermochronology, which reveal a major denudational episode initiated about 50 Ma BP, indicate that significant local relief must have existed in the early Cenozoic (Gleadow and Fitzgerald, 1987; Brown et al., 1994). A total of ~5 km of denudation has occurred since this time along the Ross Sea coast, with somewhat lesser amounts inland. This phase of denudation could have been precipitated by tectonic uplift along the rift flank, by the creation of a new, lower base level through subsidence of the Ross Embayment adjacent to a rift flank with some residual elevation, or by a combination of these two factors (Brown et al., 1994).

Although 3 km of surface uplift at a mean rate of ~1 km Ma<sup>-1</sup> since the early or middle Pliocene has been suggested for the Transantarctic Mountains (Behrendt and Cooper, 1991), compelling evidence exists from the present elevations of sub-aerially erupted volcanics (Wilch et al., 1993a), from the variation in the rate of production of cosmogenic nuclides with altitude (Brook et al., 1995; Bruno et al., 1997), and from geophysical evidence (ten Brink et al., 1993) that surface uplift in the Dry Valleys region over the past few million years has been minimal, with *maximum* possible uplift over the past ~2.5 Ma being limited to only ~300 m.

## 2.2. *Geology and morphology of the Dry Valleys area*

The Dry Valleys region represents a 4000 km<sup>2</sup> ice-free area of the Transantarctic Mountains from 77°15'S to 77°45'S and 160°E to 164°E, located between the Ross Sea embayment and the Taylor Dome on the flank of the East Antarctic Ice Sheet (Fig. 1). Along the length of the Transantarctic Mountains, the plateau periphery is cut by numerous major transverse valleys carved by outlet glaciers draining from the Polar Plateau formed by the East Antarctic Ice Sheet. Along the western coast of

McMurdo Sound, three of these valleys—the Taylor, Wright and Victoria systems—are currently largely ice-free as a result of being starved of flow from the Polar Plateau through the influence of the Taylor Dome lying immediately inland. Known collectively as the Dry Valleys, these valley systems are separated from each other by the 1500–2400 m high Asgard and Olympus Ranges.

Bedrock exposures in the Dry Valleys area consist of a basement complex of Precambrian igneous and meta-igneous rocks overlain by Devonian-to-Triassic-age sandstones, siltstones and conglomerates of the Beacon Supergroup, which dip gently inland away from the Ross Sea coast. Highly localized Cenozoic volcanics and associated intrusions occur throughout the area, whereas Jurassic dolerites (Ferrar dolerites) extensively intrude the basement and overlying sedimentary sequence forming sills up to several hundred metres thick.

Several landscape elements can be distinguished in the Dry Valleys region (Sugden et al., 1995a). Here we focus on the two areally most significant components, namely: (1) the high-elevation, low-relief surfaces that generally lie above about 1800 m and are the main landform in the interior sector of the Dry Valleys area; and (2) the rectilinear slopes that flank these upland surfaces and also constitute the predominant landform element of the Dry Valleys. The most extensive low-relief surface elements occur in the high terrain between 2000 and 2400 m which overlooks the heads of the Dry Valley systems and runs along high-level interfluvies such as that formed by the Asgard Range between Wright and Taylor Valleys. These surfaces rest on near-horizontal dolerite sills and Beacon Supergroup sedimentary rocks and are, at least to some extent, structurally controlled. Sufficiently prominent to be commented upon by the early explorers of the region (Taylor, 1914), rectilinear slopes at a typical angle of 33–37° characterize the flanks of the high surfaces fringing the Polar Plateau and a significant proportion of the valley sides of the Dry Valleys. Across the higher, western part of the region rectilinear slopes generally occur below free faces with an angle of more than 60° and lead down into a colluvium-mantled footslope at an angle of around 18° (Selby, 1971). At lower elevations to the east, free faces are less common and rectilinear slopes of adjacent valleys

meet to form sharp-edged interfluves. The rectilinear slopes are cut in bedrock and have been recognized as Richter denudational slopes (Selby, 1971, 1974, 1993). They are partly covered by a thin veneer of generally coarse rock debris which appears not to exceed a thickness of 1 m even at the base of the rectilinear slope segment. Some boulders show evidence of weathering and degradation through surficial staining by iron oxide and the development of cavernous weathering forms and tafoni.

### 2.3. Climate of the Dry Valleys area

The Dry Valleys area presently experiences a hyper-arid, polar climate. A mean annual temperature of  $-19.8^{\circ}\text{C}$  at an elevation of 123 m in Wright Valley is indicative of conditions at low elevations in the bottom of the Dry Valleys (Schwerdtfeger, 1984). However, mid-summer temperatures exceeding  $+5^{\circ}\text{C}$  over several days have been recorded in middle Wright Valley (Bull, 1966) and the distribution of debris flows, mudflows, and channels demonstrates that liquid water is common below  $\sim 800$  m under the present climate. Assuming a lapse rate of  $10^{\circ}\text{C km}^{-1}$ , mean annual temperatures at higher elevations in the Asgard and Olympus Ranges, the Quatermain Mountains and along the high-elevation surfaces fringing the Polar Plateau drop to  $-30^{\circ}\text{C}$  to  $-40^{\circ}\text{C}$ , and mean annual precipitation is  $< 10$  mm water equivalent (Schwerdtfeger, 1984; Fortuin and Oerlemans, 1990). Easterly winds carry moist air with a relative humidity of around 65–75% inland from the Ross Sea. Snow precipitation reaches a mean maximum of around 100 mm per annum at the

eastern ends of the Dry Valleys. Strong katabatic winds with a generally low relative humidity of 5–60% drain from the Polar Plateau through the Dry Valleys towards the Ross Sea, although in the summer they are largely confined to the western part of the area and to the valley bottoms. The accumulation of wind-blown snow in the lee of topographic barriers sustains small, cold-based glaciers on the sides of Taylor, Wright and Victoria Valleys. Meltwater seems to be absent at high altitudes inland, although it is present at progressively higher elevations towards the coast. This trend probably results from a combination of the maritime influence near the coast and more active katabatic winds in the upper parts of the Dry Valleys (Marchant and Denton, 1996).

### 3. Characteristics of sample sites

Sample sites were chosen to assess rates of denudation in the contrasting morphological settings of high-elevation surfaces and rectilinear slopes (Table 1). In selecting specific sample locations, particular attention was paid to identifying representative sites from which useful extrapolations about rates of denudation could be made and that would yield the maximum geomorphologically useful information. Although in sampling from discrete bedrock outcrops the possibility always exists of recording atypical rates of denudation, the landform elements from which we sampled are relatively homogeneous over lateral distances of hundreds of meters and, thus, some spatial extrapolation of site-specific data is

Table 1  
Sample and field data

Sample no.	Area	Location	Elevation	Lithology	Surface slope	Exposure geometry
50/95	Arena Valley	S 77°46', E 160°51'	1572 m	Arena sandstone	2°	Partially shielded
51/95	Arena Valley	S 77°46', E 160°51'	1572 m	Arena sandstone	0°	Partially shielded
52/95	Arena Valley	S 77°46', S 77°51'	1572 m	Arena sandstone	38°	Partially shielded
56/95	Beacon/Arena interfluve	S 77°53', E 160°44'	2050 m	Beacon sandstone	0°	Partially shielded
57/95	Beacon/Arena interfluve	S 77°53', E 160°44'	2050 m	Beacon sandstone	0°	Partially shielded
5/96	Mount Fleming	S 77°33', E 160°08'	2427 m	Beacon sandstone	0°	100% exposure
6/96	Mount Fleming	S 77°32', E 160°16'	2038 m	Beacon sandstone	0°	100% exposure
13/96	Lower Taylor Valley	S 77°43', E 162°38'	810 m	Granite	25°	Partially shielded

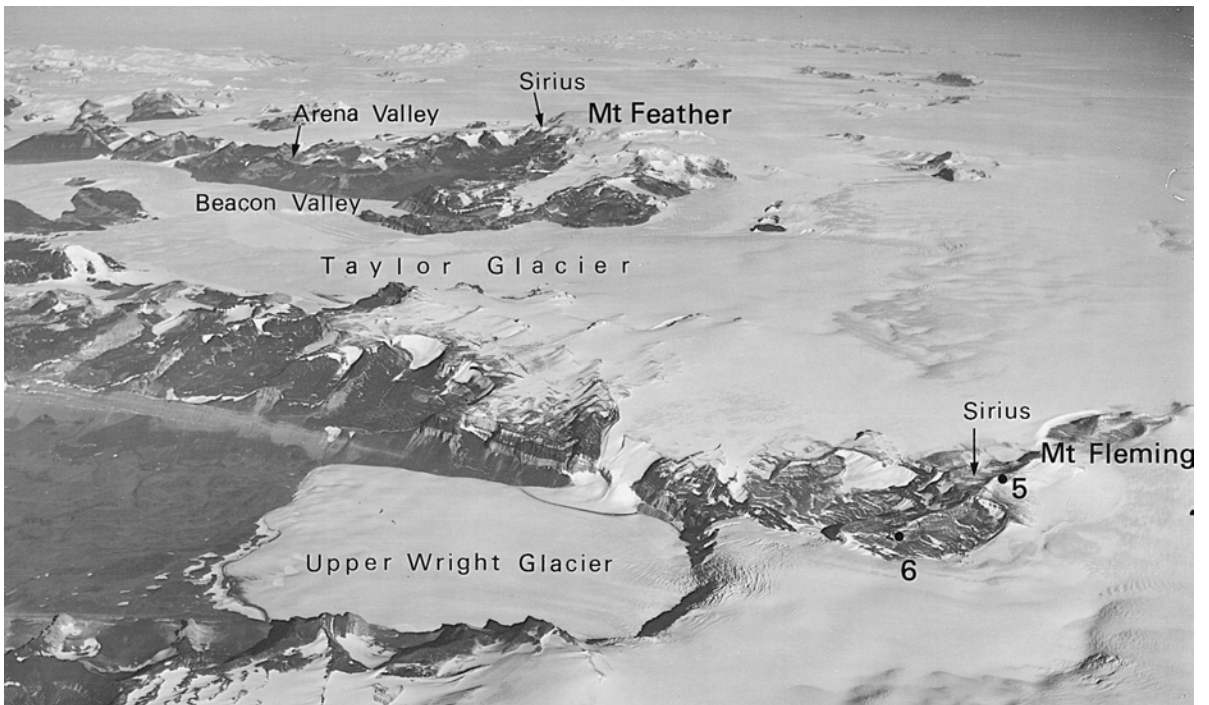


Fig. 2. Oblique air photo showing the location of sampling sites 5/96 and 6/96 on the east flank of Mount Fleming at the head of Wright Valley. Sampling sites 50–52/95 and 56–57/95 are in Arena Valley. Locations of the glaciogenic Sirius Group deposits are also shown.



Fig. 3. Sample site 5/96 on the east flank of Mount Fleming at an elevation of  $\sim 2400$  m. Note the thin snow cover at this highly exposed location. Wright Upper Glacier can be seen in the middle distance.

justified. We sampled bedrock outcrops in preference to regolith or colluvium as the exposure history of the latter under the highly episodic transportational regime operating in the Dry Valleys is much more uncertain. Sampling sites on rectilinear slopes were chosen to provide comparisons with existing independent evidence of rates of landscape change from dated lava and ash deposits (Wilch et al., 1993b; Marchant et al., 1996). Our sampling strategy, therefore, differed significantly from previous studies of cosmogenic isotopes in Antarctic samples which focused on the dating of particular deposits or exposure events rather than characterizing variations in rates of modification of the landscape. Sample locations were recorded using GPS equipment and elevations were determined to  $\pm 20$  m using altimeters calibrated daily to the elevation of the field camp in lower Taylor Valley.

Four samples were analyzed from high-elevation (> 2000 m), low-relief surfaces, all consisting of

Beacon Sandstone (Beacon Supergroup). Two of these were collected from the eastern flank of Mount Fleming at the head of Wright Valley overlooking Wright Upper Glacier (Fig. 2), one on a broad, low-relief ridge just below the summit (5/96) (Fig. 3), and the other from a lower elevation on a flat-topped spur from an iron-stained bedrock surface scattered with dolerite boulders and cobbles (6/96). The other two high-elevation samples (56/95 and 57/95) were collected from the flat-topped inter-fluve between Beacon Valley and Arena Valley above Taylor Glacier. Bedrock surfaces at all of the high-elevation surface sampling sites exhibit evidence of some rock disintegration in the form of cavernous weathering and the development of tafoni. A thin (< 100 mm), discontinuous snow cover was observed at the higher sample site at Mount Fleming (5/96), although significant shielding by snow of the bedrock surface from exposure to cosmic radiation on the high-elevation surfaces is unlikely be-



Fig. 4. Rectilinear slopes on the western side of Arena Valley, Quatermain Mountains. Free faces are evident above the rectilinear segment which terminates upslope in the dark dolerite sill. The location of samples 50–52/95 is marked by a cross.

cause the very high wind speeds, characteristic of these extremely exposed locations, retard snow accumulation.

Analyses were undertaken of three samples (50–52/95) of Arena Sandstone (Beacon Supergroup) from within a 10 m<sup>2</sup> area on a 36–38° rectilinear slope in the Quartermain Mountains on the western side of Arena Valley overlooking Taylor Glacier. The samples were collected at an elevation of ~1570 m, on a rectilinear slope ~500 m below the high-elevation surface sample site (56–57/95) located on the adjacent interfluvium (Fig. 4). Near-horizontal sandstone beds protrude from the slope and form a series of cavernously weathered steps with localized undercutting and slumping. Sample 50/95 was taken from one of the bedrock steps which characterize the slope and which have a relief of up to 4 m. Sample 51/95 is from a ~0.4 m boulder on the same bedrock step, whereas sample 52/95 was collected from the slumped front of the step. The site was selected because of its proximity to an avalanche deposit (one of two identified on the slope) located 100 m to the south which contains <sup>40</sup>Ar/<sup>39</sup>Ar-dated (11.3–12.9 Ma BP) volcanic ash, and which had

been previously used to infer minimal rates of landscape change in this area over the past several million years (Marchant et al., 1993b, 1996). This ash-avalanche deposit extends two-thirds of the way down the rectilinear slope from a bedrock couloir and comprises sandstone and dolerite clasts, quartz sand and dolerite grus with minor granite erratics chaotically mixed with a phonolitic ash incorporating glass shards and euhedral anorthoclase crystals. A sharp contact occurs between the debris cover of the rectilinear slope and the avalanche deposit which rises up to ~3 m above the surrounding talus mantle. Additional evidence of minimal geomorphic activity on the slope is provided by a thinly distributed granite till located only 30 m downslope from the sample site for 50–52/95 which has been dated by its association with volcanic ash to >13.6 Ma BP (Denton et al., 1993).

Analysis was undertaken of one further sample (13/96) collected from a north-facing ~33° rectilinear slope in lower Taylor Valley immediately to the east of Sollas Glacier (Fig. 5). The slope is formed in granite basement and is characterized by protruding bedrock outcrops with a discontinuous



Fig. 5. Site of sample 13/96 immediately to the east of Sollas Glacier, Taylor Valley. The sample site is indicated by the silhouetted figure on the left and the view is to the west across the Sollas Glacier and along Taylor Valley towards the Quartermain Mountains and Arena Valley (sample site 50–52/95). The dated lava-flow (arrowed) can be seen on the slope to the right resting on the lighter granitic bedrock. No evidence exists that this originally extended as far as the sample site.

mantle of coarse talus. The sample comprised fragments from a  $\sim 10$  mm thick exfoliating sheet of granite on an in situ bedrock exposure inclined at  $\sim 25^\circ$ . The site was chosen because it is immediately adjacent to a  $^{40}\text{Ar}/^{39}\text{Ar}$ -dated (2.2 Ma BP) partially eroded pyroclastic and lava-flow deposit (Wilch et al., 1993b). This site is at a sufficiently low elevation to have been temporarily covered by expansion of Taylor Glacier during the Quaternary.

#### 4. Analytical methods

The samples were crushed and sieved, and the 250–500  $\mu\text{m}$  fractions were selected for analysis. Mono-mineralic quartz was prepared by magnetic separation and selective chemical dissolution following the procedure described by Kohl and Nishiizumi (1992). Aliquots were set aside for future  $^{10}\text{Be}$  and  $^{26}\text{Al}$  analysis. The process of dissolution etches up to 15  $\mu\text{m}$  from the surface of the quartz crystals and this effectively removes a large proportion of implanted alpha particles produced in adjacent minerals. Samples were then ultrasonically cleaned in deionised water to remove traces of HF from mineral surfaces. Two samples (5/96 and 6/96) contained a small proportion of darker grains (possibly those with zircon inclusions) and these were removed by hand picking. Approximately 250 mg of each sample was wrapped in aluminium foil and loaded into the extraction system and evacuated to  $< 10^{-8}$  torr for 48 hours prior to analysis.

Noble gases were extracted by heating samples for 15 minutes to  $1400^\circ\text{C}$  in a double vacuum resistance furnace with a tungsten heating element and a molybdenum crucible. The extracted gas was cleaned on two Ti-getters (at  $250^\circ\text{C}$  and  $800^\circ\text{C}$ ) and a SAES-getter (at room temperature). The heavy noble gases (Ar, Kr and Xe) were separated from He and Ne by successive absorption on two charcoal traps cooled with liquid nitrogen. Neon was then absorbed on to a charcoal trap at  $-228^\circ\text{C}$  and He isotope determinations were made on the residual gas. Subsequently, Ne was released from the charcoal at  $-173^\circ\text{C}$  and the isotopic composition was analysed after the He isotope measurements. All measurements were made on a VG 5400 noble gas mass

spectrometer. This has an ion source with a modified Nier-type geometry and is equipped with an axial electron multiplier and an off-axial Faraday cup. The amplifiers of the multiplier and the Faraday cup have switchable resistors (multiplier  $10^8$  and  $10^9$ ; Faraday cup  $10^9$ ,  $10^{10}$  and  $10^{11}$ ).

The abundances of noble gas were determined by peak height comparison with known amounts of gas.  $^3\text{He}$ -enriched geothermal gas ( $^3\text{He}/^4\text{He} = 14.3 \pm 0.1 R_a$ , where  $R_a$  is the atmospheric ratio of  $1.39 \times 10^{-6}$ ) was used for the He calibrations. The He elemental and isotopic abundances in the geothermal standard were determined by repeated cross-calibrations with 0.25  $\text{cm}^3$  STP air. Neon calibrations were made on  $95.2 \pm 0.5 \mu\text{cm}^3$  STP air. The reproducibility of noble gas abundances was better than  $\pm 1.5\%$  and isotopic ratios of replicate calibrations were better than 0.5%.  $^{20}\text{Ne}$  and  $^{22}\text{Ne}$  were corrected for interfering  $^{40}\text{Ar}^{2+}$  and  $\text{CO}_2^{2+}$ , respectively, using values of  $^{40}\text{Ar}^{2+}/^{40}\text{Ar}^+$  and  $\text{CO}_2^{2+}/\text{CO}^{2+}$  determined previously. The abundances of  $\text{H}_2$  and  $\text{H}_2^{16}\text{O}$  were routinely measured, but displayed no significant variation through the week of analysis. Blanks at  $1400^\circ\text{C}$  were  $^4\text{He}$ :  $2 \times 10^{-11} \text{cm}^3$  STP and  $^{20}\text{Ne}$   $1 \times 10^{-11} \text{cm}^3$  STP.  $^3\text{He}$  was undetectable in the blank and the Ne blanks were indistinguishable from the atmospheric isotopic composition.

#### 5. Results and interpretation

The isotopic composition of Ne and He, and the concentrations of  $^{21}\text{Ne}$  and  $^3\text{He}$  are presented in Table 2. In all cases, the measured  $^{21}\text{Ne}/^{20}\text{Ne}$  (0.00489–0.0183) are significantly greater than the atmospheric value (0.00296). At its simplest, the  $^{21}\text{Ne}/^{20}\text{Ne}$  of minerals are a mixture of atmospheric and cosmogenic contributions. Nuclear reactions on  $^{18}\text{O}$ , however, produce nucleogenic  $^{21}\text{Ne}$  ( $^{21}\text{Ne}_n$ ) throughout the lifetime of the rock which may obscure the simple two component mixing. The concentration of radiogenic  $^4\text{He}$  ( $^4\text{He}_{\text{rad}}$ ) of a sample is a monitor of the  $\alpha$ -particle flux (Niedermaier et al., 1993) which is essential for  $^{21}\text{Ne}_n$  production (Wetherill, 1954). Thus, the  $^{21}\text{Ne}_n$  contribution may be estimated from the  $^4\text{He}_{\text{rad}}$  content using the most recent determination of the  $^{21}\text{Ne}_n/^4\text{He}_{\text{rad}}$  of crustal

Table 2  
He and Ne isotope data for Antarctic quartz samples

Sample	Weight (mg)	$^4\text{He}$ $10^{12}$ atoms $\text{g}^{-1}$	$R/R_a$	$^3\text{He}_c$ $10^8$ atoms $\text{g}^{-1}$	$^{20}\text{Ne}$ $10^{10}$ atoms $\text{g}^{-1}$	$^{21}\text{Ne}/^{20}\text{Ne}$ ( $\times 10^{-3}$ )	$^{21}\text{Ne}_c$ $10^8$ atoms $\text{g}^{-1}$
<i>Rectilinear slopes</i>							
50/95	263	$9.80 \pm 0.01$	$9.39 \pm 0.1$	$1.28 \pm 0.02$	$0.08 \pm 0.001$	$6.25 \pm 0.06$	$0.88 \pm 0.01$
51/95	257	$107.1 \pm 0.3$	$1.58 \pm 0.1$	$2.36 \pm 0.03$	$3.27 \pm 0.01$	$10.41 \pm 0.04$	$1.51 \pm 0.03$
52/95	320	$9.77 \pm 0.08$	$4.69 \pm 0.06$	$0.64 \pm 0.02$	$3.83 \pm 0.02$	$8.37 \pm 0.04$	$2.11 \pm 0.02$
13/96	250	$2.35 \pm 0.01$	$0.65 \pm 0.05$	$0.005 \pm 0.001$	$1.31 \pm 0.01$	$5.11 \pm 0.06$	$0.38 \pm 0.01$
<i>Plateau surfaces</i>							
56/95	302	$11.14 \pm 0.03$	$25.1 \pm 0.3$	$4.12 \pm 0.45$	$3.45 \pm 0.01$	$18.78 \pm 0.06$	$5.65 \pm 0.05$
57/95	263	$20.67 \pm 0.04$	$1.75 \pm 0.02$	$0.50 \pm 0.01$	$3.19 \pm 0.01$	$18.97 \pm 0.08$	$5.22 \pm 0.05$
5/96	278	$9.05 \pm 0.01$	$32.78 \pm 0.7$	$4.13 \pm 0.09$	$7.39 \pm 0.02$	$11.27 \pm 0.06$	$6.19 \pm 0.04$
6/96	288	$46.6 \pm 0.02$	$11.94 \pm 0.14$	$7.75 \pm 0.09$	$6.29 \pm 0.02$	$10.46 \pm 0.06$	$4.51 \pm 0.04$

$^3\text{He}/^4\text{He}$  ( $R$ ) are expressed relative to the air ratio ( $R_a$ )  $1.4 \times 10^{-6}$ . See text for method of calculation of  $^{21}\text{Ne}_c$  and  $^3\text{He}$ . The quoted errors are those of individual measurements at the  $1\sigma$  level. Abundance measurements have a reproducibility of  $\pm 1.5\%$ .

rocks ( $4.5 \times 10^{-8}$ ; Yatsevich and Honda, 1997). Assuming that all the measured  $^4\text{He}$  is radiogenic, and that at least 5% has been retained since cooling below the  $^{21}\text{Ne}_n$  closure temperature, then the contribution of  $^{21}\text{Ne}_n$  is less than 5% in all but one sample (51/95 = 30%). This can be considered as a conservative estimate because, in all but one sample, more than 5% of the  $^3\text{He}_c$  has been retained despite it being significantly more mobile in quartz than the  $^4\text{He}_{\text{rad}}$  (Trull et al., 1991).

The  $^{21}\text{Ne}_c$  concentrations displayed in Table 2 have been calculated from the measured values of  $^{21}\text{Ne}$  assuming an atmospheric composition and corrected for the contribution of nucleogenic  $^{21}\text{Ne}$ .  $^3\text{He}/^4\text{He}$  are in the range 0.65–32.8  $R_a$  (Table 2). These ratios are significantly in excess of radiogenic values and they identify the presence of high cosmogenic  $^3\text{He}$  concentrations ( $^3\text{He}_c$ ). The  $^3\text{He}_c$  concentrations have been calculated assuming that the inherited  $^3\text{He}$  is radiogenic ( $^3\text{He}/^4\text{He} = 0.02 R_a$ ).  $^3\text{He}_c$  concentrations are insensitive to the precise value used for the radiogenic  $^3\text{He}/^4\text{He}$  because of the high ratio of the samples. Errors in  $^{21}\text{Ne}_c$  and  $^3\text{He}_c$  abundances are propagated from the analytical errors and do not include uncertainties in the reproducibility of abundance measurements ( $\pm 1.5\%$ ).

The minimum apparent exposure ages of all samples can be calculated from the measured  $^{21}\text{Ne}_c$  and  $^3\text{He}_c$  abundances assuming no denudation (Table 3).  $^{21}\text{Ne}_c$  and  $^3\text{He}_c$  rates of production in quartz have not been measured directly but estimated by calibra-

tion against cosmogenic  $^{26}\text{Al}$  abundances. This procedure yields rates of production of 21 atoms  $^{21}\text{Ne}_c$   $\text{g}^{-1} \text{a}^{-1}$  (Niederman et al., 1994) and 85 atoms  $^3\text{He}_c$   $\text{g}^{-1} \text{a}^{-1}$  at sea level at high latitude (Brown et al., 1992). The rates of production have been adjusted to the altitudes of our sample sites using the nuclear disintegration data of Lal (1991). Corrections were also made to the rates of production for the effect of topographic shielding and the dip of the sampled surface. These were calculated on the basis of the

Table 3  
Data for He and Ne isotope age of exposure and rate of denudation for Antarctic quartz samples

Sample	$^3\text{He}_c$ age (ka)	$^{21}\text{Ne}_c$ age (ma)	$^{21}\text{Ne}_c$ denudation rate ( $\text{m Ma}^{-1}$ )
<i>Rectilinear slopes</i>			
50/95	$373 \pm 4$	$1.04 \pm 0.01$	$0.6 \pm 0.006$
51/95	$687 \pm 6$	$1.79 \pm 0.03$	$0.35 \pm 0.002$
52/95	$186 \pm 2$	$2.48 \pm 0.02$	$0.26 \pm 0.02$
13/96	$11 \pm 3$	$0.61 \pm 0.006$	$1.02 \pm 0.01$
<i>Plateau surfaces</i>			
56/95	$791 \pm 9$	$4.66 \pm 0.04$	$0.133 \pm 0.001$
57/95	$102 \pm 1$	$4.30 \pm 0.04$	$0.142 \pm 0.002$
5/96	$647 \pm 14$	$3.90 \pm 0.01$	$0.159 \pm 0.003$
6/96	$1,594 \pm 19$	$3.78 \pm 0.01$	$0.164 \pm 0.001$

Errors in exposure ages and erosion rates are errors propagated from experimental measurements. They do not include those from uncertainties from reproducibility of abundance measurements ( $\pm 1.5\%$ ) and in production rates ( $\pm 20\%$ ; Niederman et al., 1994).

angular dependence of cosmic radiation,  $F(\theta) = \sin^{2.3}\theta$ , and in all cases were less than 3%. The resulting  $^{21}\text{Ne}_c$  minimum apparent exposure ages range from 610 ka on a rectilinear slope (13/96) to 4.66 Ma on a high-elevation surface.

The  $^3\text{He}_c$  apparent exposure ages are between 2 and 42 times lower than those derived from the  $^{21}\text{Ne}_c$  abundances (Table 3) (Fig. 6) because of preferential diffusive loss of  $^3\text{He}_c$ . The relative age differences are unrelated to apparent exposure age, lithology, altitude or temperature. The absence of any systematic response clearly confirms the limitations of obtaining accurate exposure chronologies from  $^3\text{He}_c$  in quartz (e.g., Graf et al., 1991).  $^3\text{He}_c$  exposure ages of coarse quartz chips from the Antarctic which have not been treated by the chemical cleaning procedure used here are significantly closer to the  $^{21}\text{Ne}_c$  age (Dunai and van der Wateren, unpublished data). This suggests that laboratory handling procedures may promote the diffusive loss of  $^3\text{He}_c$  in quartz.

Concentrations of stable cosmogenic nuclides reach an equilibrium state with respect to production and loss through denudation assuming a finite, constant rate of denudation and continuous, long-term exposure (Lal, 1995). Although we cannot fully verify these two assumptions, field evidence indicates that the predominant form of denudation on the high-elevation surfaces is through a continuous incremental detachment of clasts or thin slivers of

rock, with some wind abrasion. These conditions also satisfy the assumption that the increments of denudation are very small in relation to the mean length of cosmic-ray absorption (attenuation length) of  $\sim 0.5$  m ( $\sim 150$  g  $\text{cm}^{-2}$ ) (Lal, 1991). On the rectilinear slopes, the detachment of thicker rock masses is evident in the case of the Arena Valley site, although the granitic basement at the lower Taylor Valley site is exfoliating in sheets  $< 20$  mm thick at the point where it was sampled. For reasons discussed below, the assumption of long-term exposure of bedrock surfaces at our sample sites is supported by independent geochronological and stratigraphic data. We, therefore, consider it more appropriate to interpret these  $^{21}\text{Ne}_c$  concentrations as representing long-term rates of denudation rather than apparent exposure ages (Table 3). Although we regard the possibility as being excluded by existing cosmogenic isotope data (Brook et al., 1995; Bruno et al., 1997) as well as independent geochronology from volcanic deposits (Wilch et al., 1993a), any significant surface uplift at our sample sites during exposure would cause an over-estimation of rates of denudation because the  $^{21}\text{Ne}_c$  rate of production increases with altitude.

Modifying the equation given by Kurz (1986) for stable cosmogenic nuclides in rock surfaces which have been exposed for short times relative to the formation age of the rock (clearly the case for our samples), we can interpret the measured concentration ( $N$ , atoms  $\text{g}^{-1}$ ) of a stable cosmogenic isotope,  $i$ , in terms of a maximum constant rate of denudation ( $\varepsilon$ ,  $\text{g cm}^{-2} \text{ a}^{-1}$ ), such that:

$$\varepsilon = (P \times l) / N_i \quad (1)$$

where  $P$  is the rate of production (atoms  $\text{g}^{-1} \text{ a}^{-1}$ ) and  $l$  is the cosmic-ray attenuation length ( $\text{g cm}^{-2}$ ). To convert rates of denudation into  $\text{cm a}^{-1}$  for a specific site, the result can be divided by the density of the sample ( $\text{g cm}^{-3}$ ). The rates of denudation displayed in Table 3 have been calculated using a  $^{21}\text{Ne}_c$  attenuation length of  $165$   $\text{g cm}^{-2}$  (Sarda et al., 1993), and a mean rock density of  $2.7$   $\text{g cm}^{-3}$ . All rates reported here have been converted to  $\text{m Ma}^{-1}$ .

The highest rate of denudation of  $1.02$   $\text{m Ma}^{-1}$  is for sample 13/96 from a rectilinear slope in lower Taylor Valley, although this is probably an overesti-

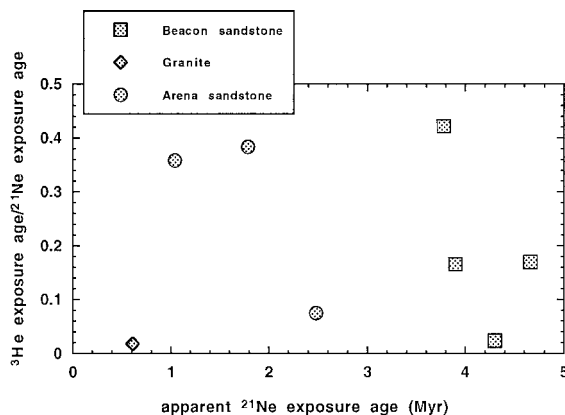


Fig. 6.  $^{21}\text{Ne}$  apparent exposure age for the analyzed samples plotted against the ratio of  $^3\text{He}/^{21}\text{Ne}$  age illustrating the variable diffusive loss of  $^3\text{He}$ .

mate given the possibility of temporary burial of this site under an expanded Taylor Glacier during the Quaternary. All four samples from the high-elevation surfaces above 2000 m display much lower rates of 0.133–0.164 m Ma<sup>-1</sup>. Samples from the rectilinear slope on the western side of Arena Valley (50–52/95) yielded intermediate rates of denudation of 0.26–0.60 m Ma<sup>-1</sup>. Assuming prolonged steady-state denudation, and given the residence time of the samples in the cosmic-ray attenuation zone (~2 m depth), these rates apply over time scales of > 1 Ma for the highest rate of 1.02 m Ma<sup>-1</sup>, and > 4 Ma for the lowest rate of 0.133 m Ma<sup>-1</sup>.

## 6. Discussion

Our data elucidate various aspects of the morphological evolution of the Dry Valleys area of the Transantarctic Mountains and the late Cenozoic history of the East Antarctic Ice Sheet. The issues we consider here are: (1) the interpretation of the measured <sup>21</sup>Ne<sub>c</sub> concentrations as indicative of rates of denudation rather than exposure ages; (2) the observed variations in rates of denudation between sample sites and the interpretation of these in terms of weathering and erosional processes, and the magnitude of late Cenozoic modification of the landscape; (3) the implications of the <sup>21</sup>Ne<sub>c</sub> data for existing interpretations of late Cenozoic lavas and volcanic ash deposits; (4) the comparison of rates of denudation reported here with previous estimates for the Dry Valleys and rates of denudation for contrasting morphoclimatic regimes; and (5) the implications of our data for the ice-sheet stability debate.

### 6.1. *Geomorphological interpretation of cosmogenic <sup>21</sup>Ne data*

When expressed as apparent ages of surface exposure (Table 3), the older ages reported here tend to be associated with the higher elevation sites. Old apparent ages of exposure for high-altitude, inland locations compared with somewhat younger ages at lower elevations towards the coast have been noted elsewhere in the Transantarctic Mountains (Nishizumi et al., 1991; Giegengack et al., 1994). A

potential explanation for this trend is that lowering of the East Antarctic Ice Sheet progressively exposed bedrock surfaces at lower elevations. Compelling arguments, however, exist against this mechanism as a general explanation of the data. For instance, climatic reasons exist for expecting a gradient of decreasing weathering and rates of erosion inland from the coast, and from low to higher elevations, in the Dry Valleys sector of the Transantarctic Mountains which could account for the trend observed in apparent ages of exposure. In addition, independent paleoceanographic, geochronometric, stratigraphic and geomorphological evidence for the history of the East Antarctic Ice Sheet (Denton et al., 1993; Kennett and Hodell, 1993) indicates stability since the Miocene and argues against the <sup>21</sup>Ne<sub>c</sub> data being interpreted as a record of the timing of progressive bedrock exposure from beneath a wasting ice mass. Apart from a modest expansion of glaciers in the Pliocene, Marchant et al. (1993b) have argued that no significant change has occurred in the extent of glaciers lying above an elevation of 1500 m in the western Dry Valleys area, since the deposition of a granite-rich drift by East Antarctic outlet glaciers more than 13.6 Ma ago. This interpretation has been further supported by the discovery of a preserved remnant of granite-bearing glacier ice at the mouth of Beacon Valley overlooking Taylor Glacier (Sugden et al., 1995b). The ice has a minimum age of 8 Ma BP on the basis of <sup>40</sup>Ar/<sup>39</sup>Ar dating of overlying in situ volcanic ash. Any significant thickening of Taylor Glacier at the mouth of the valley, or of the East Antarctic Ice Sheet, would have caused this remnant ice to have been overridden and removed by normal processes of ice deformation. The ice preserves detailed characteristics from the time of initial deposition and demonstrates that no such overriding occurred.

### 6.2. *Denudational processes and degree of late Cenozoic modification of the landscape*

Although our data indicate very low rates of denudation for all the sample locations, a somewhat higher rate of denudation of around 1 m Ma<sup>-1</sup> is indicated for the site at low elevation in lower Taylor Valley (13/96). Little can be inferred solely from

this single sample, but the partially eroded form of the immediately adjacent 2.2 Ma old lava-flow (Wilch et al., 1993a,b) supports the interpretation of a higher rate of geomorphic activity in this area. Indeed, the occurrence of gelifluction lobes, patterned ground, debris flows, and ephemeral stream channels in the lower Taylor Valley indicates some level of geomorphic activity reflecting the presence of meltwater, relatively high precipitation and humidity, and freeze–thaw cycles during the summer. With the absence of liquid water further inland and at higher elevation, however, the landscape of the western Dry Valleys is essentially relict (Marchant and Denton, 1996). The low rates of denudation reported here for the rectilinear slope in Arena Valley (samples 50–52/95) and the even lower rates of  $< 0.2 \text{ m Ma}^{-1}$  on the high-elevation surfaces (samples 5–6/96 and 56–57/95), probably reflect the exceedingly slow operation of salt weathering in conjunction with deflation, and very rare mass movements precipitated by rock failures on rectilinear slopes and free faces (Augustinus and Selby, 1990; Selby, 1971, 1974). The relative abundance of tafoni and cavernous weathering forms suggests that salt weathering is more active at lower elevations near the coast where salt supply and humidity are at a maximum. More uncertain is the role of cryptoendolithic microorganisms which can currently survive in the Dry Valleys up to an elevation of about 2000 m in the vicinity of Mount Fleming (Friedmann et al., 1994). They may promote case hardening of bedrock surfaces through silicification as well as rock breakdown through biochemical processes (Weed and Norton, 1991). Irrespective of the precise role of these various weathering and erosional processes, they vary in efficacy across the Dry Valleys area. Rates of denudation vary in consequence even under the present climatic regime.

Amounts of denudation since the most recent episode of wet-based glacial erosion of the Dry Valleys have been sufficient to allow differential rates of weathering and erosion of contrasting lithologies to produce relief of a few metres. As noted by Selby (1974), fine-grained rocks, such as lamprophyre and finely crystalline dolerites, form positive relief forms where they intrude less resistant granites. Our  $^{21}\text{Ne}_c$  data, in combination with the survival of dated volcanic deposits, indicate, how-

ever, that a maximum of only a few metres of total denudation has occurred throughout most of the Dry Valleys during the past several million years. Therefore, the presently active geomorphic processes have led to only minimal modification of the landscape in these areas during this extended period of time and have not been capable of creating the present topography from an earlier landscape shaped by through-valley glaciation.

These extremely slow rates of landscape change of  $\sim 0.2\text{--}1.0 \text{ m Ma}^{-1}$  can be compared with estimates of longer term rates of denudation from fission-track thermochronology. In the Dry Valleys area, clear evidence exists for a period of accelerated rates of denudation initiated in the early Cenozoic (50–55 Ma BP) associated with rifting, with a less certain phase of Cretaceous denudation being identified inland (Gleadow and Fitzgerald, 1987; Fitzgerald, 1992). The amount of denudation since the early Cenozoic is at a maximum of  $\sim 5 \text{ km}$  in the vicinity of the rift–flank axis and Ross Sea coastal zone, and declines inland to the west. This gives a maximum mean rate of denudation for the past  $\sim 50 \text{ Ma}$  of  $\sim 100 \text{ m Ma}^{-1}$ , although thermal modelling of fission-track age and track length data indicate that rates of crustal cooling (denudation) were higher ( $\sim 200 \text{ m Ma}^{-1}$ ) for a period of about 10–15 Ma after the initiation of this late Cenozoic phase of accelerated denudation. Although such fission-track data do not provide information on more recent denudation of the upper  $\sim 1 \text{ km}$  of the crust, the reduction in rates of denudation, evident after the initial pulse associated with rifting  $\sim 55 \text{ Ma}$  ago, is fully compatible with these  $^{21}\text{Ne}_c$  results and other geochronological data that demonstrate extremely low rates of denudation over the past several million years.

### 6.3. Correlations with lava and volcanic ash deposits

The  $\sim 1 \text{ m Ma}^{-1}$  rate of denudation recorded for the slope site in lower Taylor Valley (13/96) is fully consistent with the partially eroded nature of the immediately adjacent 2.2 Ma old lava-flow (Fig. 5) (Wilch et al., 1993a,b). Potentially more equivocal, however, is the interpretation of the ash-avalanche deposits on the slopes of Arena Valley as truly in

situ, especially given the late Miocene age ascribed to these sediments. These deposits are of considerable importance given the significance they have in the debate about the late Cenozoic history of the East Antarctic Ice Sheet (Marchant et al., 1993b). One view might be that they represent pods of volcanic ash originally deposited within hollows on the summit surface that have subsequently avalanched down the slope when intersected by slope retreat. Another interpretation might be that they have been emplaced at some time after original eruption by aeolian processes. In both scenarios, the radiometric age of the ash would provide no information on the rate of landscape change in Arena Valley. By contrast, evidence that these deposits represent avalanching essentially contemporaneous with initial airfall includes physical characteristics of the ash that are diagnostic of primary ash falls, with limited subsequent reworking. These characteristics include poor sorting (bi-modal grain-size distribution of glass shards), intact bubble vesicles, and angular glass shards. That each ash-avalanche deposit is composed of material from a single eruption is also supported by a distinct geochemistry and the consistency in radiometric age of individual crystals (Marchant et al., 1996). It has, therefore, been inferred that these ash-avalanche deposits formed shortly after times of eruption, with strong katabatic winds deflating ash exposed at the surface, but being unable to detach material incorporated within the slope-mantling talus (Marchant et al., 1996).

Our  $^{21}\text{Ne}_c$  data for the slope site in Arena Valley (samples 50–52/95), located close to the 11.3–12.9 Ma dated ash-avalanche, support the interpretation of its formation essentially contemporaneously with initial deposition. If extrapolated, the estimated denudation rates of  $\sim 0.2\text{--}0.6\text{ m Ma}^{-1}$  imply minimal slope retreat of only about 3–4 m over the past 11 Ma under the present hyper-arid, polar climatic regime. These would be sufficiently low to preserve the ash-avalanche deposit. Such low rates of denudation would also be compatible with preservation without significant disturbance of moraine ridges of Pliocene age which cross the lower parts of the avalanche cone. Volcanic ash might be expected to survive such limited slope retreat through progressive infiltration into voids created by mechanical weathering of the slope-mantling talus and underlying

ing bedrock. Survival would also be expected if the original avalanche deposit was originally more than 3–4 m thick.

#### 6.4. Comparison of rates of denudation

Our maximum denudation rates of  $< 1\text{ m Ma}^{-1}$  are within the range previously reported for studies of cosmogenic isotopes from sites in the Dry Valleys sector and adjacent areas of the Transantarctic Mountains. For instance, Nishiizumi et al. (1991) have modelled maximum rates of denudation based on concentrations of cosmogenic  $^{10}\text{Be}$  and  $^{26}\text{Al}$  for nine samples from the Allan Nunatak (Allan Hills), 75 km northwest of Victoria Valley. These samples range in age from  $0.24\text{ m Ma}^{-1}$  to  $1.31\text{ m Ma}^{-1}$  (mean =  $0.65\text{ m Ma}^{-1}$ ). In the same study, four samples from Wright Valley gave a range of  $0.54\text{ m Ma}^{-1}$  to  $1.31\text{ m Ma}^{-1}$  (mean =  $0.93\text{ m Ma}^{-1}$ ). Similarly, low rates of  $0.06\text{--}0.27\text{ m Ma}^{-1}$  over the past 2–3 Ma, based on concentrations of cosmogenic  $^{10}\text{Be}$ , have been recorded by Brook et al. (1995) for quartz arenite boulders and cobbles in glacial deposits in the Quartermain Mountains and Asgard Range in the Dry Valleys area. Extremely low maximum rates of denudation, based on concentrations of cosmogenic  $^{26}\text{Al}$  and  $^{10}\text{Be}$  and ranging up to  $0.7\text{ m Ma}^{-1}$ , have also been reported by Ivy-Ochs et al. (1995) for Beacon Supergroup sandstones, granite from glaciogenic Sirius Group deposits from Table Mountain at the head of Ferrar Glacier, and sandstone boulders from a Sirius Group deposit at Mount Fleming. Rates of denudation of less than  $0.1\text{ m Ma}^{-1}$  have also been reported by Bruno et al. (1997) from concentrations of cosmogenic  $^{21}\text{Ne}$  for pyroxene, quartz and whole-rock dolerite samples from Sirius Group tillites and associated bedrock samples from the Dry Valleys.

Lack of detailed information on the geomorphic setting of the sites sampled in these studies makes direct comparison with our own results difficult, but clearly the low overall rates of denudation reported are comparable with the estimates from the  $^{21}\text{Ne}_c$  data reported here. Together with existing results, our data support the contention that the Dry Valleys sector of the Transantarctic Mountains have the lowest recorded rates of denudation for any terrestrial environment, especially in the case of the high-eleva-

tion surfaces. They only appear to be matched by the maximum rates of denudation of as low as  $0.6 \text{ m Ma}^{-1}$  reported by Bierman and Turner (1995) on the basis of concentrations of  $^{26}\text{Al}$  and  $^{10}\text{Be}$  in samples from granitic domes in the Eyre Peninsula, South Australia. By contrast, much higher rates of denudation of  $5\text{--}11 \text{ m Ma}^{-1}$  have been derived from concentrations of cosmogenic  $^{26}\text{Al}$  and  $^{10}\text{Be}$  for rhyolitic volcanic ash-flow tuffs in New Mexico, USA (Albrecht et al., 1993) and of  $\sim 3.2 \text{ m Ma}^{-1}$  for a flat interfluvium in Réunion (Sarda et al., 1993). Mean rates of denudation for large drainage basins estimated from data for modern sediment and solute load range from  $5 \text{ m Ma}^{-1}$  to  $688 \text{ m Ma}^{-1}$  and show that rates for the Dry Valleys area lie well below the average for low relief basins (Summerfield and Hulton, 1994). Mean rates of denudation for individual catchments may be as low as  $1 \text{ m Ma}^{-1}$  (Summerfield, 1991), but these figures include large areas of net deposition, and, thus, underestimate local rates of denudation comparable to the site-specific rates provided by the analysis of cosmogenic isotopes.

#### *6.5. Implications for the debate about ice-sheet stability*

By supporting the in situ interpretation of the ash-avalanche deposits in Arena Valley, and by demonstrating the extremely low rates of denudation for various sites in the Dry Valleys area, our  $^{21}\text{Ne}_c$  data strongly support the scenario that the East Antarctic Ice Sheet has been essentially stable since the mid-Miocene (Marchant et al., 1993a,b; Sugden et al., 1995a,b). Our  $^{21}\text{Ne}_c$  data certainly seem incompatible with the opposing view of ice-sheet instability and major deglaciation during the Pliocene associated with much warmer conditions (Webb et al., 1984; Barrett et al., 1992) which would have promoted significantly higher rates of denudation than is indicated by the concentrations of  $^{21}\text{Ne}_c$  reported here. Our results, and interpretation of long-term climatic stability, are also in accord with other studies using cosmogenic isotopes (Brook et al., 1995; Ivy-Ochs et al., 1995) that demonstrate a minimum late Miocene age for key Sirius Group deposits that represent the most recent significant ingress of glacial ice into the Dry Valleys area.

## 7. Conclusions

Our strategically selected samples from contrasting components of the landscape in the Dry Valleys area yield concentrations of cosmogenic  $^{21}\text{Ne}$  that indicate extremely low rates of denudation of  $\sim 0.15 \text{ m Ma}^{-1}$  on low relief, high-elevation surfaces and  $\sim 0.20\text{--}1.00 \text{ m Ma}^{-1}$  on rectilinear slopes. Field evidence and independent geochronological data support the interpretation of concentrations of  $^{21}\text{Ne}_c$  as reflecting rates of denudation rather than ages since exposure from beneath an ice cover. Our estimated rates are comparable to those established from concentrations of cosmogenic isotopes for other locations in the Dry Valleys and confirm that denudation rates in this hyper-arid polar environment are the lowest of any terrestrial environment, and are around two orders of magnitude lower than mean rates over the past  $\sim 50 \text{ Ma}$  estimated from fission-track thermochronology. Such slow rates of geomorphic activity, even on high angle rectilinear slopes, largely involving salt weathering, deflation, and limited mass movement, mean that only minimal modification of the landscape has occurred in the Dry Valleys area since the most recent phase of extensive glaciation around  $15 \text{ Ma}$  ago. This interpretation is supported by the survival of volcanic deposits extending back to the mid-Miocene, and our estimated rates of denudation for the slopes in Arena Valley are compatible with the in situ survival of ash-avalanches dating back to more than  $11 \text{ Ma BP}$ . Our data are at variance with the hypothesis that the Dry Valleys were overrun by glacial ice following a major deglaciation event as recently as the late Pliocene; rather, when considered in conjunction with related field and geochronometric data, our results support the view that the East Antarctic Ice Sheet has been essentially stable for at least the past  $15 \text{ Ma}$ .

## Acknowledgements

This research was supported by the UK Natural Environment Research Council [grant no. GR3/9128 (DES/MAS) and research studentship no. GT4/93/8/G (HAPC)]. Field support was provided

by the Division of Polar Programs of the National Science Foundation of the USA through grants to the University of Maine. Alun Hubbard provided assistance with the exposure geometry corrections to production rates.

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