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Estimates of late Holocene soil production and erosion in the Snowy Mountains, Australia

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Abstract

Soil production in actively uplifting or high precipitation alpine landscapes is potentially rapid. However, these same landscapes are also susceptible to erosion and can be sensitive to changes in climate and anthropogenic activity which can upset the balance between soil production and erosion. The Snowy Mountains, southeastern Australia, are a tectonically stable, low relief, moderate precipitation mountain environment. The alpine area is extensively blanketed by soil that has been subjected to more intensive episodes of erosion during past periods of anthropogenic disturbance and under cold climate conditions of the late Quaternary. In this study, rates of soil development and hillslope erosion were investigated using radiocarbon dating, fallout radionuclides and sediment cores collected from lakes and reservoirs. Estimated Holocene soil development rates were 20-220 t/km2/y. Erosion rates determined from the radionuclides 137Cs and 210Pb were equivocal, due to the inherent spatial variability of radionuclide inventories relative to apparent erosion rates. Estimated average erosion rates over the past 100 years, determined from 210Pbex inventories, were 60 t/km2/y (95% CI: 10, 90). Inventories of 137Cs observed at the same site implied that more recent erosion rates (over the past 60 years) was below the detection limits of the sampling method applied here (i.e. /km2/y). The upper estimate of 90 t/km2/y is comparable to the mean erosion rate estimated using the radionuclide method for uncultivated sites in Australia and is significantly lower than that measured at sites were vegetation cover was disturbed by livestock grazing prior to its exclusion from the alpine area in the 1940s CE. Low erosion and high soil production rates relative to the lowland soils are likely related to extensive vegetation cover, which, in this context, protects soils against erosion and contributes to the formation of organic alpine soils, that rapidly accumulate organic matter by comparison to other soil types.

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Estimates of late Holocene soil production and erosion in the Snowy Mountains, Australia

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Abstract

Soil production in actively uplifting or high precipitation alpine landscapes is potentially rapid. However, these same landscapes are also susceptible to erosion and can be sensitive to changes in climate and anthropogenic activity which can upset the balance between soil production and erosion. The Snowy Mountains, southeastern Australia, are a tectonically stable, low relief, moderate precipitation mountain environment. The alpine area is extensively blanketed by soil that has been subjected to more intensive episodes of erosion during past periods of anthropogenic disturbance and under cold climate conditions of the late Quaternary. In this study, rates of soil development and hillslope erosion were investigated using radiocarbon dating, fallout radionuclides and sediment cores collected from lakes and reservoirs. Estimated Holocene soil development rates were 20-220 t/km^2/y. Erosion rates determined from the radionuclides ^{137}Cs and ^{210}Pb were equivocal, due to the inherent spatial variability of radionuclide inventories relative to apparent erosion rates. Estimated average erosion rates over the past 100 years, determined from $^{210}Pb_{ex}$ inventories, were 60 t/km^2/y (95% CI:10, 90). Inventories of ^{137}Cs observed at the same site implied that more recent erosion rates (over the past 60 years) was below the detection limits of the sampling method applied here (i.e. <70 t/km^2/y). The upper estimate of 90 t/km^2/y is comparable to the mean erosion rate estimated using the radionuclide method for uncultivated sites in Australia and is significantly lower than that measured at sites were vegetation cover was disturbed by livestock grazing prior to its exclusion from the alpine area in the 1940s CE. Low erosion and high soil production rates relative to the lowland soils are likely related to extensive vegetation cover which in this context protects soils against erosion and contributes to the formation of organic alpine soils, which rapidly accumulate organic matter by comparison to other soil types.

Keywords: erosion; soil production; Holocene; anthropogenic impacts; climate; alpine

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1. Introduction

On a global scale, alpine landscapes are recognised as regions of relatively high geomorphic activity (Dedkov and Moszherin, 1992; Milliman and Syvitski, 1992). This has been ascribed to their climate, which is typically cold and wet, tectonic activity and corresponding high elevations and steep slopes (high potential energy), which in combination promote rapid physical weathering, erosion and sediment transport (Milliman and Syvitski, 1992; Syvitski and Milliman, 2007; Vanmaercke et al., 2011; Walling and Webb, 1996). For example rivers draining mountain basins transport a disproportionately large proportion of the global sediment yield, that is, 870 t/km^2/y compared to 115 t/km^2/y for the rest of the world's rivers (Milliman and Farnsworth, 2011). Especially in tectonically active mountain ranges, erosion can be so rapid as to equal or exceed the rate of uplift (Brozović et al., 1997; Koppes and Montgomery, 2009; Mitchell and Montgomery, 2006).

Despite high rates of geomorphic activity, the historical perception was that in cold mountain environments, rates of chemical weathering and, therefore, soil formation were low (e.g. Peltier, 1950). Contrary to this assumption, rapid rates of soil production have been measured in alpine environments, especially in regions experiencing rapid uplift and high precipitation. For example, in the Southern Alps New Zealand, where rainfall may exceed 10 m/y and uplift approximates 10 mm/y, soil production rates may reach 2.5 mm/y (Larsen et al., 2014), an order of magnitude higher than soil production rates measured elsewhere (Larsen et al., 2014). The potential significance of chemical weathering in mountain environments is further evidenced by the existence of extensive soil mantles in a variety of alpine settings worldwide (Dixon and Thorn, 2005; Egli et al., 2014; Norton and von Blanckenburg, 2010; Riebe et al., 2004).

Soils in high mountain environments have been shown to be sensitive to changes in climate and human activity, which alter soil production processes and may accelerate the erosion rates by multiple orders of magnitude (Barsch and Caine, 1984). Episodic changes in climate, vegetation cover, fire frequency and human disturbance are, therefore, likely to be important controls on the balance between soil production and erosion in some mountain environments (Hewawasam et al., 2003; Kirchner et al., 2001; Koppes and Montgomery, 2009; Schmidt et al., 2002).

In tectonically stable, non-glaciated, high to moderate rainfall, i.e. less than 2 m/y, alpine areas, such as the Snowy Mountains in south-eastern Australia, rates of soil production and erosion have received less attention. The Snowy Mountains have traditionally been viewed as distinct from other alpine regions (Costin, 1989; Kirkpatrick, 1994). This is due to their intra-plate setting and resulting tectonic stability and moderate relief (slopes) (Bishop and Goldrick, 2000). In addition, they experienced relatively limited Pleistocene glaciation (Barrows, 2001). These characteristics have facilitated the development of a relatively thick soil mantle (0.6-> 1 m) over almost the entire alpine area (Costin, 1989). Nevertheless, the Snowy Mountains are considered to have experienced pulses of intensified sediment transport in response to changing climate during the late Quaternary (Costin, 1972; Kemp and Rhodes, 2010; Ogden et al., 2001; Page et al., 2009) and as a result of livestock grazing between the mid-1800s and 1940s CE (Costin et al., 1960).

The quantification of inherently spatially and temporally variable soil production and erosion rates remains a major challenge within geomorphology. Measurement of hillslope erosion has been undertaken by a variety of methods that can be broadly categorised into: plot and survey approaches (e.g. Costin et al., 1960); measurement of sediment yields by either stream gauging, or by measurement of the mass of sediment accumulated in geomorphic sinks, such as lakes (e.g. Neil, 1991; Tomkins et al., 2007); erosion tracer methods using fallout radionuclides (e.g. ^{137}Cs and ^{210}Pb) (Blake et al., 2009; Loughran et al., 1988; Porto et al., 2009; Ritchie and McHenry, 1990; Walling et al., 2003) and; the application of cosmogenic nuclides (e.g. Dixon and Riebe, 2014; Heimsath et al., 2002). These methods are each limited by the challenges of upscaling point measurements in time and space in relation to the representativeness of reference sites, the spatial heterogeneity of tracer fallout and transport and the issues of sediment storage and delivery (Chappell et al., 2011b; de Vente et al., 2007; Zhang et al., 2015). In addition, these methods provide data over different time periods, e.g. stream gauging typically provides short term data (event to decadal scale), radionuclides provide decadal to centennial scale data, and commonly used cosmogenic nuclides integrate over millennial scales. As a result, different approaches will commonly yield very different results (e.g. Tomkins et al., 2007; Wasson et al., 1996) that are then subject to various interpretations.

The objective of this study is to quantify soil development and erosion rates in a tectonically stable, currently non-glaciated mountain environment and to advance the understanding of the relative controls that changing climate and anthropogenic activities place on landscape stability and sediment budgets. In doing so the likely age of these soils is discussed, which in this setting is likely to be constrained by glaciation and/or periglacial processes to at least <11-16 ka (Barrows, et al., 2001, Costin, 1972). This study employs multiple methods to attempt to quantify rates of soil development, hillslope erosion and sediment transport. Hillslope erosion rates are investigated using fallout radionuclides (^{137}Cs and $^{210}Pb_{ex}$) and by calculating sediment mass accumulation rates in alpine lakes and reservoirs. Soil development rates are examined using geomorphic and paleoclimate evidence combined with radiocarbon analysis. These approaches overlap in time, allowing the balance between soil development and erosion rates to be investigated. Results are placed within the context of historical data quantifying erosion under past conditions of anthropogenic and climatic change.

2. Regional setting

The Snowy Mountains are a high elevation plateau of moderate (undulating) relief. Despite being the highest region of Australia, they reach only 2228 m at their highest point (Mt Kosciusko) and local relief of the alpine area is usually less than 200 m. The Snowy Mountains are the erosional remnants of uplift associated with the Cretacous breakup of Gondawana beginning 100 ma with most intense activity centered around 55 ma (Bishop and Goldrick, 2000). Their intraplate setting results in low denudation rates (2-5 m/Ma) over geological timescales ((Young and McDougall, 1993). The basement rocks of the mountains are Silurian and Devonian aged granites with Ordovician meta-sediments and occasional Tertiary basalts. The Snowy Mountains contain the only peaks above 2,000 m in Australia and form part of mainland Australia's limited subalpine and alpine area, which covers only 2500 km^2.

Aligned perpendicular to the prevailing westerly moisture bearing winds, the Snowy Mountains experience a cool montane climate, with mean temperature varying from 18° C in summer to -7° C in winter with annual precipitation ~ 2000 mm (BOM, 2014) . In the alpine tract, continuous snow cover is present for up to 4 months of the year with isolated snow patches sometimes persisting through the year (Green and Pickering, 2009). Interannual variability of snow-depth and snow persistence is high and is related to the frequency of occurrence of snow bearing synoptic weather systems (Nicholls, 2005; Theobald et al., 2015; Whetton et al., 1996). Minor periglacial activity (needle ice formation and frost heave) is currently confined to the alpine zone, with gelifluction and frost shattering most significant above 2000 m (Barrows et al., 2001; Galloway, 1965).

In the montane zone (900-1500 m), vegetation comprises wet-sclerophyll forests dominated by alpine ash (*Eucalyptus delgatensis*) and mountain gum (*E. dalrympleana*). Subalpine (above 1500 m) vegetation consists largely of snowgum (*E. pauciflora*) woodlands with a grassy (*Poa ceaspitossa*) understory. Above 1850 m are alpine herbfields (dominated by *Celmisia and Poa spp*) with areas of heath, sod tussock grassland and fen-bog communities (*Carex-Sphagnum*). Vegetation cover is almost complete, extending to the highest peaks.

From the mid 1800's to the mid 1900's CE the Snowy Mountains were exploited for summer grazing of sheep and cattle (Green et al., 2006). By the late 1920s CE, deliberate burning by graziers and trampling of vegetation by stock had resulted in the initiation of soil erosion over a substantial area of the alpine zone (Bryant, 1971). Surveys undertaken at this time implied that sheet erosion was the predominant erosion process (Bryant, 1971), however gully erosion also occurred particularly around the summits of the highest peaks (Irwin and Rogers, 1986). Since 1944 CE, when it was declared a National Park the entire alpine area of the Snowy Mountains has been protected from grazing and resource extraction.

Soils in the alpine and subalpine zone of the Snowy Mountains are classed as humose chernic tenosols (Australian Soil Classification) (McKenzie et al. 2014). At the study location, they are characterised by a humose A horizon of approximately 30 cm depth with an abrupt to distinct transition to a relatively shallow stony BC horizon. The BC horizon grades to the granite substrate at approximately 60 cm depth. The soils are considered polygenetic (Brewer and Haldane, 1972). The A horizon has been assumed to be Holocene in age and is dominated by high organic inputs from the covering snowgrass accompanied by slow decomposition due cold temperatures, frequently waterlogged soils and high aluminium content (McKenzie et al., 2004). The soils are augmented by significant dust accretion (Costin et al., 1952; Johnston, 2001; Marx et al., 2011). Together this suggests that the A horizon undergoes pedogenisis by upbuilding. This differs markedly from the underlying poorly developed B/C horizon which has been considered to represent a truncated Pleistocene palaeosol developed during a period of greater chemical weathering (which were originally termed pedoderms; Brewer and Haldane, 1972). The presence of stone lines below the A horizon, as occurs commonly in the Snowy Mountains, has been argued to represent an erosional feature consistent with the polygenetic origin of the soils (Brewer and Haldane, 1972).

The onset of soil production of the chernic tenosols in the alpine region of the Snowy Mountains has been assumed to date from sometime after the end of glaciation (< 15ka) (Barrows, et al., 2001; Brewer 1972, McKenzie, 1994). Moraines dating from 15.5 ka occur immediately below one of the study sites (Cootaptamba) (Barrows, 2001), while significant periglacial activity has been assumed to have resulted stripping of soil alpine soils elsewhere in the Snowy Mountains(Costin, 1972). Development of alpine peat mires have variously dated between to 7-18 ka (Costin, 1972; Martin, 1999; Marx et al., 2011), implying the onset of significant organic production at these times and also potentially constraining the onset of humose soil formation.

3. Methods

3.1 Study locations

Estimates of rates of soil development, hillslope erosion and sediment yield were undertaken within the catchment of Guthega Reservoir, located in the headwaters of the Snowy River (Fig. 1). Guthega catchment includes the highest peaks of the Main Divide. Sixty-five percent of the catchment lies within the alpine zone, which is characterised by both high relative precipitation and high runoff coefficients (Reinfelds et al., 2014). Much of this area is comprised of high elevation plateaux surfaces or low gradient valley floors and, as a result, the mean catchment slope is only 12°. Only 20% of the catchment area has a gradient greater than 20°, distinguishing it from many other alpine settings. The topography of Guthega catchment is representative of the Kosciusko alpine region in general, with the exception of the steeper western facing slopes of the range and the valleys of the major rivers.

Estimates of erosion (via ^{137}Cs and $^{210}Pb_{ex}$) and soil production were made on a hill slope in the sub-catchment of Guthega Creek (-36.367°, 148.368°, Fig. 1), which drains directly into Guthega Reservoir. The slope has a westerly aspect with an elevation range of 1880 m AHD at the ridge crest to 1670 m at Guthega Creek. The local treeline is located at approximately 1800 m, just below the ridge crest. Snowgrass (*Poa spp)* predominates above the treeline, with open snowgum (*Eucalyptus pauciflora*), with a dense grassy understorey (*Poa and Chionachloa frigida* (Ribbony Grass)) below. Together these provide close to 100% vegetation cover. This site is considered broadly representative of alpine and subalpine zones in the Snowy Mountains in terms of soil and vegetation and slope characteristics. Mean slope at the site is 16°, compared to a median of 12° for the Snowy Mountains as a whole.

Erosion measurement by ^{137}Cs and $^{210}Pb_{ex}$ was also undertaken across a second hillslope at Cootapatamba, which lies within the alpine zone, at the ridgeline of the Main Range. Site elevation is 2160 m AHD at the ridge-crest and 1910 m AHD at the valley floor. The hill slope has a northwesterly aspect. It is densely covered by alpine herbfield vegetation with a mean slope of 14°.

Catchment sediment yield was estimated from sediment mass accumulation rates in cores collected from Guthega Reservoir (-36.379, 148.371, 1578 m) and Club Lake (-36.414, 148.291 1950 m), a small cirque lake located within the headwaters of the reservoir catchment. Guthega Reservoir has an area of 0.26 km^2 with a contributing catchment area of 91 km^2. The area of Club Lake area is 0.015 km^2 with a contributing catchment area of 0.3 km^2. Catchment soils are largely stabilised by alpine herbfield vegetation, although evidence of past instability is provided by gravel deposits buried within fen peats on the margins of the lake (Dodson et al., 1994). The phyllite bedrock is exposed on the steeper sections of the cirque walls and areas of scree are also present. The mean cirque catchment slope is 22° and 20 % of the catchment has a gradient greater than 32°. There is therefore likely higher potential for erosion in the Club Lake catchment compared with the rest of the Snowy Mountains region due to its steeper slopes and exposed sediment.

3.2 Measurement of soil development rates

A potential range for the rate of soil development was estimated by two separate methods. In the first approach, the likely date of onset for the Holocene stage of soil development was ascertained from geomorphic evidence and previous studies dating the reestablishment of stable conditions following the last ice age in southern Australia. Soil development rates were then estimated by dividing the mass of accumulated soil by its likely age. A second estimate was made from the radiocarbon ages of soil organic matter (SOM) at multiple depths within soil profiles at Guthega (Wang et al., 2014). Samples from each of three soil profiles from the Guthega hillslope were dated by ^{14}C AMS and the mass of soil accumulated between each dated interval was determined from the product of the increment volume and the corresponding dry bulk density. These approaches were used in combination to provide a potential range rather than absolute estimate of the rate of soil development.

In this context, it is acknowledged that radiocarbon ages are typically younger than the true soil age (Wang et al., 1996). This method has received critical discussion within the radiocarbon literature, the chief identified limitation being the contaminating effect of new carbon which is incorporated into the profile during pedogenesis (Goh et al., 1977; Scharpenseel and Becker-Heidmann, 1992; Wang et al., 1996). The soils formed on the hillslopes of Snowy Mountains alpine zone, however, display properties which suggest that they may build upwards in a manner similar to peat, that is, dead organic material accumulates at the top of the profile (further discussion of which is provided in the section 2 of this manuscript). Accordingly, it was considered that the variation in radiocarbon content through the depth of the soil profile could provide insight into relative rates of soil formation, within the acknowledged limitations of this method. Radiocarbon measurement was performed on the humin SOM fraction, which has been argued to be the environmentally stable SOM component and, therefore, relatively less susceptible to contamination by newly incorporated carbon (Kristiansen et al., 2003; Pessenda et al., 2001; Wang et al., 2014). Radiocarbon ages undertaken on soil humins have been shown to reasonably estimate true age in a range of modern soils (Leavitt et al., 2007; Pessenda et al., 2001).

For the current study, samples were collected from soil profiles excavated at three locations along a transect (catena) down the slope. These were the ridge crest (-36.360, 148.366, 1866 m), the mid-slope (-36.360, 148.370, 1804 m) and the toe-slope (-36.360, 148.366, 1665 m). Profiles were excavated until the underlying fractured bedrock or saprolite was reached (approximately 0.6 m), with samples collected at approximately 100 mm depth intervals. The deepest sample dated, in each case, came from the base of the A horizon at its intersection with the B/C horizon, i.e. the presumed stripped Pleistocene surface (Fig. 2).

Sample pre-treatment and analysis was performed by ^{14}C Accelerator Mass Spectrometry (AMS) at the Waikato Radiocarbon Dating Laboratory and at the UC Irvine Keck-CCAMS facility (UCI KCCAMS), respectively. To remove visible organic contaminants (modern roots), soil was first disaggregated by stirring with sodium pyrophosphate on a hotplate and then was washed through a 1 mm sieve. The <1 mm fraction was retained and evaporated to dryness. The soil was then lightly homogenised in a mortar and pestle. Humins were separated from more mobile fractions, such as fulvic and humic acids, prior to graphitisation using the acid-base-acid method, according to the UCI KCCAMS Facility Acid/Base/Acid sample pre-treatment protocol. Samples were washed first in hot HCL then with NaOH to remove soil humics. The NaOH insoluble fraction was washed again in hot HCL, then rinsed and dried. Radiocarbon ages were calibrated using the Oxcal programme (v. 4.2) (Bronk-Ramsey, 2009) and the SHCal13 calibration curve (Hogg et al., 2013).

Soil organic content was estimated by dry combustion. Sub-samples from each soil depth interval were oven dried then ashed at 450° C for 12 hours.

3.3 Estimating hillslope erosion rates

Quantification of hillslope erosion was undertaken using the radionuclide tracers $^{210}Pb_{ex}$ and ^{137}Cs, where $^{210}Pb_{ex}$ (half life 22.3 yrs) measures erosion rates over time scales of approximately 100 years and ^{137}Cs measures erosion rates since approximately 1960 CE (i.e. following significant ^{137}Cs fallout from atomic testing).This method models soil loss or gain based on the measured deviation of the radionuclide inventory from the local reference inventory, which is determined at a non-eroding site (Walling and He, 1999). Quantitative estimates of erosion or deposition rates can be derived from the relationship between the magnitude of deviation and the extent of soil loss or gain, for which a number of empirical (e.g. Loughran et al., 1988; Ritchie and McHenry, 1975) or process based (Walling and He, 1999) models have been developed.

Soil cores were collected from three transects at Guthega hillslope and three transects at Cootapatamba hillslope (Fig. 1). Six cores were collected from each transect (18 cores in total) using 100 mm diameter PVC tube. Soil was collected to a depth of 15 cm or until the stone line, a feature of some Snowy Mountains soils (Brewer and Haldane, 1972), prevented further excavation. Fifteen additional cores were collected from non-eroding reference site(s) which were located at the ridge crests of the respective hillslopes. Samples were oven dried to constant weight, homogenised, and sieved through a 2 mm sieve. The <2 mm fraction was ground in a rock mill to a fine powder, packed into petri dishes and sealed for approximately three weeks to achieve secular equilibrium between ^{222}Rn and its daughter products, including ^{214}Pb. Samples were analysed for ^{210}Pb and ^{137}Cs by gamma spectrometry at the Institute for Environmental Research, Australian Nuclear Science and Technology Organisation (ANSTO). The activity of $^{210}Pb_{ex}$ was calculated as the difference between total measured ^{210}Pb and ^{226}Ra activity (which was derived from the activity of its granddaughter isotope ^{214}Pb). Count times were between 1 and 5 days, depending on the total sample activity, resulting in mean counting errors of 6 ± 0.4 % for ^{137}Cs. Mean $^{210}Pb_{ex}$ counting errors were 13 ±2 %. Activity of ^{137}Cs was decay corrected to February 2013 and all $^{210}Pb_{ex}$ activities were decay corrected to their sampling date. Radionuclide inventories for individual cores were calculated by multiplying the measured radionuclide activity of the core by its mass and then dividing by its surface area.

To account for the intrinsic variability of radionuclide inventories and the resulting uncertainty of erosion estimates produced using the available conversion models, this study adopted the recommendations of Zhang (2014) and Kirchner (2013). These authors argue that the usual approach used when modelling erosion volumes from radionuclide inventories, i.e. comparing individual points at the suspected eroding site to single value representing the reference inventory (usually the reference site mean), is problematic in that it ignores the contribution of random spatial variability and sampling error to the deviation of any point measurement from the reference mean (Kirchner, 2013; Zhang et al., 2015).

In reality, the reference mean is unlikely to accurately represent the original inventory at the point in question (Wallbrink and Murray, 1996). For this reason, prior to modelling being undertaken, statistical anlaysis (t-test, independent samples median test and independent samples Kolmogorov-Smirmov test) was undertaken to establish whether the observed differences between the reference and hillslope inventories were greater than that which would be expected due to random spatial variability and sampling error. Net erosion, or deposition, was assumed to have occurred only if the deviation from the reference inventory exceeded that expected due to random spatial variability and sampling error at the 95% confidence level ($p<0.05$). The best estimate of the percentage change in radionuclide inventory, representing the volume of soil lost or gained from the hillslope, was taken as the difference between the mean of each hillslope and that of its respective reference site. An estimate of the potential range of actual erosion rates was obtained by rerunning the conversion models using the upper and lower 95 % confidence limits of both the reference site and hillslope means.

The ^{137}Cs and $^{210}Pb_{ex}$ inventories were converted to erosion volumes using the theoretical diffusion and migration models of Walling et al. (2007), using their software. The equations for these models are provided in Walling et al. (2007). In the case of ^{137}Cs inventories an alternative model, the Australian Empirical Model (AEM) (Elliott et al., 1990; Loughran et al., 2004) was also used. The equations for this model are provided in Loughran et al. (2004). The AEM was developed specifically for Australian conditions and is based on relationships established between ^{137}Cs and erosion plot data (Elliott et al., 1990; Loughran et al., 1988). Erosion rates estimated using the AEM have been shown to correspond well with erosion plots and sediment yield data in some areas (Martinez et al., 2009). Equally, the diffusion and migration model has produced more comparable estimates in other cases (Simms et al., 2008). Based on past studies undertaken within Australia it is possible that the AEM may underestimate and the diffusion and migration model overestimate true erosion rates for Australian soils (Martinez et al., 2009; Simms et al., 2008).

The diffusion and migration model used here considers the time-dependent fallout and subsequent redistribution of $^{210}Pb_{ex}$ and ^{137}Cs within the soil profile. As this can be affected by factors extraneous to soil redistribution, such as fallout history, radioactive decay, diffusion, leaching and bioturbation, the diffusion and migration model requires inputs describing the profile typology of radionuclide activity in soil unaffected by erosion or deposition (these include the diffusion parameter, downward migration rate and relaxation depth factor (which describes profile shape)). In order to characterise the change in radionuclide activity with depth, three cores from the reference site were selected and sectioned into 2 cm increments on which ^{137}Cs and $^{210}Pb_{ex}$ activity was measured. Model inputs (the diffusionparameter, downward migration rate and relaxation depth factor) were estimated from these cores using the models of Walling et al. (2007) (Table 1). A particle size correction factor of 1.6 (Blake et al., 2009) was applied to eroding points to account for the enrichment of radionuclides in eroded material due to the preferential erosion of smaller particles with higher ^{210}Pb activity.

The different fallout histories (essentially undetectable after the 1980s CE for ^{137}Cs and constant for naturally derived $^{210}Pb_{ex}$) of the two radionuclides mean that they integrate the net effect of soil redistribution processes over different times spans, that is approximately 60 years for ^{137}Cs and around 100 years for $^{210}Pb_{ex}$ (^{210}Pb half-life is 22.2 years) (Walling et al., 2007).

3.4 Estimating sedimentation rates

Sediment cores were retrieved from Guthega Reservoir (n=1) and Club Lake (n=1). Collection processing and dating of the Club Lake core was previously described in Stromsoe et al. (2013), while similarly the Guthega Reservoir core was described in Stromsoe et al. (2015). Both cores are described briefly here.

A 0.38 m core was collected from Club Lake using 70 mm diameter polyvinylchloride pipe, while a 0.27 m core was collected from Guthega Reservoir, using a gravity corer. Cores were extracted from approximately the centre of the lake and the reservoir.

Cores were sectioned into 2-5 mm slices. Samples from both the Guthega Reservoir (n=7) and Club Lake (n=9) cores were dated using ^{210}Pb at the Institute for Environmental Research, Australian Nuclear Science and Technology Organisation (ANTSO). In some cases the small mass of material in each slice necessitated that multiple samples be combined to obtain sufficient sample for dating. In addition, samples from Club Lake core were also dated by ^{14}C Accelerator Mass Spectrometry at Waikato Radiocarbon Dating Laboratory (n=4). Dating results and age model construction for the Club Lake and Guthega Reservoir cores are discussed in Stromsoe et al. (2013) and Stromsoe et al (2014), respectively.

Mass accumulation rates (*MAR*) (g/cm^2/y) were calculated from the product of the increment volume, derived from the age model and the corresponding dry bulk density, according to equation 1.

$$MAR = SAR * BD \qquad (1)$$

where *SAR* is the linear sediment accumulation rate (cm/y) and *BD* is bulk density (g/cm^3).

The *MAR* was converted to an annual sediment accumulation rate for the total reservoir area *AA* (t/y).

$$AA = MAR * Reservoir\ surface\ area \qquad (2)$$

AA was then used to estimate area specific catchment sediment yields according equation 3:

$$SSY = AA/CA \qquad (3)$$

where *SSY* is the specific sediment yield (t/km^2/y), *AA* is the annual reservoir accumulation (t/y) and *CA* is the contributing catchment area (km^2).

The Club Lake *SSY* is considered to be a relatively reliable estimate of the average erosion rate within the lake catchment. This is due to the relatively small catchment size and the limited potential for sediment storage within its catchment (although some eroded sediment is stored within talus slopes on the cirque wall and within fens at the margins of the lake; Mooney et al., 1997). In the case of Guthega Reservoir the mass of sediment stored within the lake may underestimate the average rate of erosion within the catchment due to storage of eroded sediment within the catchment and incomplete sediment trap efficiency of the reservoir. Therefore, the *SSY* for Guthega Reservoir was corrected for the proportion of the catchment estimated to be geomorphically disconnected from the reservoir (i.e. *SSY* was calculated only for the area of the catchment contributing sediment to the reservoir) and the trap efficiency of the reservoir for storing sediment. The contributing catchment area was determined according to the method of Fryirs et al. (1998). The trap efficiency of the reservoir was estimated from the Brune method (Brune, 1953) using the modified curve developed by Heinemann for small reservoirs (Heinemann, 1981) according to equation 4.

$$TE_{BR} = 100 \ x \ ((V/Q)/0.012 = 1.02)) \qquad (4)$$

where TE_{BR} is the trap efficiency, *V* is the reservoir volume at capacity (m^3) and *Q* is the mean annual inflow (m^3/y).

It is acknowledged that erosion from such a large area may differ substantially or be considerably more spatially variable than suggested by single cores extracted from a limited number of sinks. However, the specific sediment yields calculated from the Club Lake and Guthega Reservoir cores are used here to provide estimated lower limit for catchment erosion rates for comparison with the hillslope erosion rates estimated by ^{137}Cs and ^{210}Pb$_{ex}$.

5. Results

5.1 Radiocarbon ages

For the Guthega soil catena, returned 1^{4}C AMS ages for A/BC horizon transition show good agreement between sites with ages for the ridge-crest ranging from 2330-2430 y cal. BP; 2520-2750 y cal. BP for the mid-slope and, 2150-2310 y cal. BP in the toe-slope profile (Fig. 2 and Table 2). Minimum age differences are as small as 30 years between the ridge-crest and toe-slope and do not display a consistent downslope relationship. The maximum radiocarbon age of the organic A horizon is, therefore, approximately 2150-2750 y cal. BP. As the A horizon is presumed to have developed on top of the older B/C horizon (i.e. the truncated Pleistocene palaeosol) (Brewer and Haldane, 1972), this indicates the potential onset of the Holocene stage of soil development at or prior to this date. Due to the potential incorporation of new carbon at depth, this should be regarded as a minimum estimate of the true soil age (Wang et al., 1996).

For each of the three profiles, returned radiocarbon ages were progressively younger toward the top of the profile, as would be expected in upbuilding soils as organic matter is added to the surface of the profile. In addition, soils at the same depths along the catena are of similar age, e.g. ages from 100-300 mm depth in the upper two soil profiles are within ~12 % of each other, which is close to calibration error (Table 2). This implies that is there is no obvious significant carbon mixing by mass wasting/movement processes. The lack of significant recent mass wasting or is also indicated by the coherence of the A/BC horizon ages and the lack of colluvium in the toeslope (soil profiles lower in the catena were shallower). If mass wasting had had occurred, then the toeslope profile ages would be expected to be appreciably older than those of further up the catena. It should be noted that mass movement (likely soil creep or solifuction) was previously argued to be a significant process in alpine soils of the Snowy Mountains responsible for the exfoliation of mica grains in the presumed Holocene age soils and also for relic stone lines which underlie this layer (Brewer and Haldane, 1972).

5.2 Hillslope erosion rates

5.2.1 Radionuclide reference inventories

The down -profile distribution of the radionuclide activity at the reference (presumed non-eroding) site was typical of that found within undisturbed soils, with peak ^{137}Cs activity located just below the soil surface and $^{210}Pb_{ex}$ relatively more concentrated in the upper most soil (Mabit et al., 2008) (Fig. 3). Mean ^{137}Cs reference inventories were 1008 ±35 Bq/m^2 (errors are mean counting errors) at Guthega and 886 ±47 Bq/m^2 at Cootapatamba (Table 3)., while mean reference $^{210}Pb_{ex}$ inventories were 11,198 ±517 and 6966 ±882 Bq/m^2 at Guthega and Cootapatamba, respectively (Table 3).

Reference ^{137}Cs inventories measured in this study are high compared to previous Australian measurements. Previous measurements of ^{137}Cs fallout for southeastern Australia are in approximate accordance with UNSCEAR estimate of 360 Bq/m^2 for the 30 to 40°S latitude band (UNSCEAR, 2000). For example, reported reference inventory values from southeastern Australia were: 383 Bq/m^{-2} measured at Blue Gum Creek, south coast, NSW) (-34.222°S, 150.492°; MAP 840 mm; Blake et al., 2009); 540-653 Bq/m^2 in the upper Hunter Valley, central coast NSW (-32.092°, 150.117°; MAP 620 mm; Martinez et al., 2009) and; 425 Bq/m^2, the average of multiple sites in southern Australia between -30 and -40° (Chappell et al., 2011a; Elliott, 1997)) (all values are decay corrected to 2013 for comparison).

The relatively high ^{137}Cs activity measured in Snowy Mountains soils is likely explained by the area's high precipitation, as ^{137}Cs activity tends to increase with increasing rainfall (e.g Chappell et al., 2011a; Schuller et al., 2002). The ^{137}Cs inventory for the Snowy Mountains is broadly consistent with, although still higher than, by the rainfall-inventory relationship developed for Australian sites by Chappell et al. (2011a), i.e. 745 Bq/m^2. However, it is similar to the only previous measurement from a location where mean annual precipitation (MAP) approaches that of the Snowy Mountains (, i.e. 884 Bq/m^{-2} at 1800 mm/y MAP in coastal New South Wales (Chappell et al., 2011a; Elliott, 1997).

Excess ^{210}Pb reference inventories for the Snowy Mountains sites are also comparatively high. They are approximately 5 times those reported for Blue Gum Creek, i.e. 1923 Bq/m^2 (-34.222°, 150.492°; MAP 840 mm; Blake et al., 2009) and Townsville on the north-eastern Australian coast, i.e. 1600 Bq/m^2 (-19.256°S, 146.818°E, (-19.256°, 146.818°; MAP 1140 mm; Pfitzner et al., 2004). Higher inventories in the Snowy Mountains sites are likely explained by a combination of characteristics known to increase ^{210}Pb$_{ex}$ fallout, including the area's relatively high precipitation, high rates of dust deposition (Marx et al., 2011; Stromsoe et al., 2015), and relative distance from the coast, where ^{210}Pb production is low (Garcia-Orellana et al., 2006; Preiss et al., 1996)..

Spatial variability in ^{137}Cs and ^{210}Pb$_{ex}$ reference inventories can be described by the coefficient of variation (CV). The CV of the ^{137}Cs reference inventory was 21 % at Guthega and 30 % at Cootapatamba. For ^{210}Pb$_{ex}$ the reference inventory CV was 30 % at Guthega and 50 % at Cootapatamba. Radionuclide inventories tend to display high intrinsic variability due to the small-scale variability of several factors, including: initial fallout, which is influenced for example, by the deposition of dust which scavenges ^{210}Pb (Marx et al., 2005); microtopography; vegetation; as well as runoff and infiltration (Owens and Walling, 1996; Zhang, 2014). The CV of the Snowy Mountains sites is consistent with that observed by previous studies, i.e. typically ~20 % for ^{137}Cs (Kirchner, 2013; Pennock and Appleby, 2003; Sutherland and de Jong, 1990).The intrinsic spatial variability of ^{210}Pb$_{ex}$ fallout is less well known, but CVs of ~10 % (Porto et al., 2009) to 115 % (Mabit et al., 2009) have been reported. The high spatial variability of ^{210}Pb$_{ex}$ relative to ^{137}Cs in the Snowy Mountains reference sites may be partly attributable to the relative imprecision of gamma spectrometry for determining ^{210}Pb activity compared to ^{137}Cs (e.g. Shakhashiro and Mabit, 2009).

5.2.2 Hillslope radionuclide inventories

Ceasium-137 inventories at the hill sites were similar to, or only slightly higher than, that measured at the reference sites (Fig. 4A and 4C and Table 3). The mean ^{137}Cs inventory at Guthega hillslope was 1001 ±63 Bq/m^2 (cf reference inventory 1008 ±88 Bq/m^2). At Cootapatamba hillslope the mean ^{137}Cs inventory was 928 ±66 Bq/m^2 (cf reference inventory 886 ±70 Bq/m^2). In each case, statistical tests confirmed that the mean (t-test), median (independent samples median test) and distribution (Kolmogorov-Smirmov test) of the hill sites inventories are not statistically different from the reference siteat the 95% confidence level (Table 4).

In contrast to the ^{137}Cs results, the $^{210}Pb_{ex}$ hillslope inventories imply potential loss of $^{210}Pb_{ex}$ (i.e. net soil loss) at the hillslope sites (Fig. 4B and D and Table 3). The mean $^{210}Pb_{ex}$ inventory at Guthega hillslope was approximately 2620 Bq/m^2 lower than the expected natural inventory (Figure 4B and D Table 3), while the $^{210}Pb_{ex}$ inventory at Cootapatamba hillslope is 725 Bq/m^2 lower than the expected natural (reference) inventory. Differences between hill and reference inventories are statistically significant at Guthega hillslope but not at Cootapatamba (Table 4).

5.2.3 Modelled erosion rates

For the most part, there were no statistically significant differences in the radionuclide inventories to provide convincing evidence of significant erosion at the analysed hillslopes (Table 4). An exception is Guthega hillslope, where the $^{210}Pb_{ex}$ inventory was significantly lower than the reference inventory, equating to a modelled erosion (using the diffusion and migration model of Walling et al. (2007)) rate of 60 t/km^2/yover the last ~100 years. The range of potential erosion rates at Guthega hillslope, estimated from the upper and lower 95% confidence intervals of the reference and hillslope means, was 10-90 t/km^2/y. Based on a mean soil bulk density of 0.6 g/cm^3, the 60 t/km^2/y of soil loss experienced equates to a surface lowering rate of 0.1 mm/y.

Although at Cootapatamba hillslope, the observed loss of $^{210}Pb_{ex}$, was shown to be statistically non-significant, results imply some potential loss of $^{210}Pb_{ex}$. Consequently, modelled erosion rates are provided for reference. At Cootapatamba, the calculated erosion rate was 20 t/km^2/y. However, as $^{210}Pb_{ex}$ loss at this site is small relative to its natural variability, this value is uncertain. The maximum difference between the 95% confidence intervals of the reference and hillslope mean inventories at Cootapatamba hillslope imply a potential soil redistribution rate of between 110 t/km^2/y of erosion to 76 t/km^2/y of deposition. The ^{137}Cs hillslope inventories at both sites were essentially identical, or slightly higher, than their respective reference inventories. This implies either the total volume of soil lost over the past 60 years (the timespan of the ^{137}Cs method) was negligible, and/or that the sampling resolution was insufficient to capture and accurately characterise the ^{137}Cs inventory at the hillslope site.

While the chosen sample size (n=15) is considered as sufficient in much of the radionuclide literature (see for example the review by Pennock and Appleby, 2003), statistical power analysis demonstrates that the Snowy Mountains erosion rates are near the limit detectable by the statistical tests applied in this study. The smallest overall ^{137}Cs loss or gain detectable by this study is 336 Bq/m^2, or approximately 33 % of the mean reference inventory, at Guthega and 227 Bq/m^2, or 31% of the mean reference inventory, at Cootapatamba (at an 80 % probability of detection). This rate of ^{137}Cs loss converts to an erosion rate of 70 t/km^2/y., which is towards the mid-upper end of the erosion rates modelled using $^{210}Pb_{ex}$ (see section 5.2.3). The same ^{137}Cs values equate to a modelled erosion rate of 24 t/km^2/y using the AEM model.

5.3 Sedimentation rates

Sediment mass accumulation rates (MAR) calculated using equation 1 for Club Lake and Guthega Reservoir were 470 and 2610 g/cm^2/y, respectively. These values represent the mean accumulation rates since c.1920 CE for Club Lake and c.1955 CE for Guthega Reservoir. The SSY for Club Lake catchment was 23 t/km^2/y and the (uncorrected) SSY for Guthega Reservoir catchment was 6 t/km^2/y. The estimated area of Guthega Reservoir catchment contributing sediment directly to the reservoir was 73 km^3 and the trap efficiency of the reservoir was 44%. Taking these values into account, the corrected SSY for Guthega was 13 t/km^2/y.

The Club Lake SSY is considered likely to be at the upper end of sediment yields from the Snowy Mountains, due to the relatively small size, steepness and limited potential for sediment storage within its catchment,. Sediment storage, and therefore the MAR at Guthega Reservoir, may be affected by the release of sediment through the river outlet structure at the base of the dam. Accordingly, the 13 t/km^2/y sediment yield for Guthega catchment may underestimate true sediment yields. The collective sediment yields from Club Lake and Guthega Reservoir suggest that catchment erosion rates from the Snowy Mountains exceeds 13-23 t/km^2/y.

6. Discussion

6.1 The age of alpine and subalpine soils in the Snowy Mountains and its implications for sediment production

It is generally accepted that late Pleistocene sediment erosion and transport rates in the highlands of Australia were greater than present (Kemp and Rhodes, 2010; Page and Nanson, 1996; Page et al., 1991). This is demonstrated in the major rivers which drain the Snowy Mountains (the Murray and Murrumbidgee Rivers), which are known to have experienced markedly different channel forms and sediment characteristics, i.e. braided, low sinuosity channels transporting bedload by comparison to the current meandering high sinuosity suspended load channels (Page et al., 2009), implying geomorphically unstable conditions in mountain areas which were less suitable to the development of organic soils. In high altitude catchments, the change to the present regime occurred at approximately 6-10 ka and may have been driven, in part, by decreasing erosion rates in the Snowy Mountains (Ogden et al., 2001).

As previously discussed, the chernic tenosols of the Australian Alps are considered to be polygenetic (Brewer and Haldane, 1972), with the organic A horizon ascribed a Holocene age (McKenzie et al., 2004), while the B/C horizon believed to have developed during a period of greater chemical weathering and then truncated by mass movement (Brewer et al., 1970)). While a

Pleistocene origin is presumed, the age of the B/C horizon has not yet been confirmed by suitable dating methods. In contrast, the A-B/C transition, the position of the deepest dates obtained here, marks the onset of A horizon development and the establishment of conditions favouring preservation of organic matter and soil up-building, i.e. the current phase of soil development and would document the onset of an implied period of relative geomorphic stability which is considered to mark the Holocene, or in this context at least the mid to late Holocene.

The timing of late Pleistocene glaciation and also of maximum frost shattering in the Snowy Mountains is relatively well constrained (with three glacial advances occurring between approximately 15 and 35 ka; Barrows et al., 2001), however the date of renewed (post-last glacial) soil development of is comparatively unknown. The only relevant existing dates for soils in the alpine/subalpine (>1 500 m) zone come from peat mires (Costin, 1972; Kemp and Hope, 2014; Martin, 1986; Marx et al., 2011), which develop under specific topographic and hydrological conditions, and from timing of activity of solifluction terraces (Costin, 1972). The ^{14}C dates from Guthega hillslope are therefore the first ages for organic hillslope soils which blanket almost the entire subalpine/alpine area.

Radiocarbon dating of soil organic matter (SOM) is not straightforward due to the continuous incorporation of new carbon via root decay, bioturbation and translocation of mobile organic fractions (Wang et al. 1995). Consequently, macro-carbon, such as macro-charcoal, is regarded as producing more reliable ages, but was unfortunately not present in the studied soils. Humins, as were dated in this study, are considered by some to be the SOM fraction which most closely resembles the true soil age (Pessenda et al., 2001). Although paired radiocarbon dates on the humin fraction and on macro-charcoal from the same depth in soil elsewhere have returned equivalent ages (Wang et al. 2014, Pessenda et al. 2001), it is not possible to rule out the addition of new carbon from root material to the humin fraction in this study. Consequently, radiocarbon ages presented here are considered to represent minimum ages reflecting the mean residence time

(MRT) of carbon in soil, that is, the average age of all components of different ages in the soil, allowing for renewal and decomposition.

The radiocarbon dates obtained on SOM in the A horizon of these soils would be expected to be significantly less influenced by incorporation of new carbon than many other soils. In this alpine environment, low temperatures and humid conditions inhibit mineralisation of SOM, and may facilitate the formation of organic up-building soils. While the SOM reservoir may receive new carbon from root decay, the zone of active root growth and bioturbation will shift up with time, meaning that the radiocarbon content at different depths should reflect the date at which they formed to a greater extent than for more mineral soils. This is supported by the progressive increase in radiocarbon age with depth which implies that SOM is continuously buried by younger organic matter.

The ^{14}C dates from the A–B/C horizon transition for each the three profiles examined here returned ages of c.2 500 y cal. BP (Table 2). These are considerably younger than previously assumed ages of approximately 12-16 Ka (McKenzie et al., 2004) indicated by the termination of glaciation and the decline of significant frost shattering in the Snowy Mountains (Barrows et al., 2004; Barrows et al., 2001). The implication of this discrepancy is that either soil at the A-BC transition has incorporated significant new carbon, e.g. approximately 30-40 % of the ^{14}C present at the A-B/C transition has been incorporated after initial soil formation, or alternatively the organic horizon has developed since 2500 y cal. BP. In the following sections the potential ages of alpine and subalpine soils are explored using other geomorphic and palaeoclimatic evidence.

6.1.1 Maximum possible ages for the onset of soil development in the Snowy Mountains

The maximum likely date for the onset of the current phase of soil development at Guthega is determined by the timing of the end of glaciation in the late Pleistocene and the subsequent reestablishment of conditions suitable for the establishment of vegetation and development of the organic A horizons of the chernic tenosols. On the Australian mainland, the extent of late Pleistocene glaciation, even at the height of the Last Glacial Maximum (LGM) is believed to have been limited. In the Kosciusko region, permanent ice is thought to have been restricted to small cirque glaciers above 1850 m AHD, covering a maximum area of 15 km^2 (Barrows et al., 2001). However, periglacial activity was likely to have been considerably more widespread, resulting in major slope instability possibly extending to as low 600 m above current sea level (Galloway, 1965). In the Snowy Mountains, the limits of periglacial solifluction were possibly 975 m lower than today, implying summer temperatures 9° C colder than present (Galloway, 1965). In the current alpine/subalpine zone (>1500 m AHD), periglacial slope processes were considered sufficient to initiate widespread stripping of the pre-existing soil and vegetation, producing the stony rubble which today underlies much of the alpine/subalpine zone (Costin, 1972), i.e. the current B/C horizon and the stone lines. Thus the earliest time at which the organic A horizon and associated fine-grained weathering products began to accumulate on slopes above 1500 m in the Snowy Mountains must at the least post-date the latest deglaciation of the Kosciuzko Massif.

On the Australian mainland, several lines of evidence place the most recent Pleistocene glaciation at ~35-15 ka, with maximum cooling at 21 ±2 ka (Williams et al., 2009). Rivers draining the eastern highlands experienced episodic periods of increased discharge and bed load transport (based on the dimensions of palaeo channels and the sediment texture within them) centered at 30-25, 20-18 and 18-14 ka, implying increased snowpack volumes and geomorphic instability in less vegetated headwater catchments at these times (Kemp and Rhodes, 2010; Page and Nanson, 1996; Page et al., 1991). On the Kosciusko Massif, cosmogenic ^{10}Be dating of glacial moraines demonstrates that the latest (Mt Tynnam) and least extensive glacier advance occurred at 16.8 ±1.4 ka (Barrows et al., 2001). A lack of recessional moraines implies that the subsequent deglaciation was rapid (Barrows et al., 2001). Block streams and block aprons located between 1200 - 1750 m AHD in both the Snowy Mountains and Victorian Alps also appear to have become inactive after ~12-

17 ka (Barrows et al., 2004), indicating a decline in the intensity of frost shattering. Basal dates for organic sediment overlying glacial till in the region's largest cirque lake suggest local establishment (i.e. to 1890 m AHD) of vegetation by as early as 15.8 ka (De Deckker and Shakau, unpublished data cited in Barrows et al., 2001). The end of the last bedload dominated phase on the Riverine Plain, which occurred sometime before ~12 ka, also implies increasing stability for the highlands in general (Page et al., 2009). Thus the maximum probable date for the renewed onset of soil development in the Snowy Mountains region is ~14-16 ka.

At high elevations (the current subalpine and alpine zone) there is evidence to suggest that return to suitable conditions for soil development may not have occurred until several thousand years later. At Caledonia Fen (1280 m AHD), located ~170 km SE of the Guthega site, the timing of the transition from mineral to organic sedimentation implies that the establishment of high lake levels and stable catchment slopes did not occur until ~11 ka (Kershaw et al., 2007). This sedimentation change was concomitant with a dramatic increase in eucalypt abundance and the decline of cold climate vegetation (including *Asteraceae Tubiliflorea* and *Chenopodiaceae*) indicating the rise of the treeline to close, or above, 1280 m AHD at this time. This is 600 m lower than the elevation of the current tree line at Guthega hillslope. Assuming a lapse rate of 0.77° C/ 100 m and a temperature increase of 1° C for each degree of latitude, the Caledonia Fen results suggest that the 11 ka temperature at the Guthega ridge-crest was ~3.6 ^{o}C lower than present. By implication, mean daily temperatures at Guthega hillslope may have been <- 6° C in the coldest months with maximum daily temperatures remaining below 0° C for 3-4 months of the year (at present mean maximum daily temperatures rise above freezing in all months). While this temperature estimate is imprecise, it nevertheless provides insight into the likelihood of organic soil development occurring in the alpine area at 11 ka and suggests that cold climate conditions may have limited the rate of organic soil building until after 11 ka. This is supported by pollen data from Mt Kosciusko from elevations of 1955-1960 m AHD which suggest that grassland was not present in the alpine tract until 11,000 y cal.

BP, with the transition to the current alpine community occurring sometime after 9,000 (Rain , 1974) to 7,000 y cal. BP (Martin, 1986).

The development of peat mires in the alpine zone would also have been inhibited until slopes stabilised sufficiently to allow peat growth in low gradient areas to exceed its rate of removal (Hope, 2003). Early ^{14}C dates from the Snowy Mountains placed the onset of peat growth above 1800 m AHD at ~11–18 ka (Costin, 1972; Martin, 1986). More recently obtained dates, however, show that peat development at many sites did not occur until ~7-9 ka (Martin, 1999; Marx et al., 2011). This suggests that summer temperatures in the current alpine zone may have remained ~≤ 10°C until 7-8 ka, after which they warmed sufficiently to permit peat growth in topographically favourable positions, and/or to permit more extensive development (Marx et al., 2011). This approximately coincides with a decline in alpine pollen and rise in *Eucalpyptus* at Diggers Creek peat mire (at approx. 6500 cal. y BP) indicating the advance of the local treeline above ~1700 m (Martin, 1999).

The only previous estimate for the onset of hillslope soil development in the alpine zone comes from Pounds Creek (at 1960 m AHD), where soil overlies 15 cm of fine granitic gravel, dated to 9000 radiocarbon years (Costin, 1972; Martin, 1986). This was interpreted as representing a period of intense periglacial activity preceding the onset of soil development. Elsewhere, soil developed on late Pleistocene substrates in the Blue Mountains (>900 m AHD, 250 km north of Guthega) and dated by optically stimulated luminescence (OSL) shows a marked discontinuity in age with depth, related to accelerated erosion above the treeline during the LGM. OSL ages jump from 32 ka at 40 cm depth to 7 ka at 30 cm, implying sediment accumulation on top of the relict erosional surface began slightly after 8 ka (Wilkinson et al., 2005).

A delayed return to stable conditions in high elevation catchments is supported by the timing of the initiation of modern fluvial activity in the upper Murray River and its tributaries, which at 7-10 ka, post-dates that of the lower elevation Goulburn and mid Murray Basin by 1-10 ka (Ogden et al., 2001; Page et al., 2009; Pels, 1969). This period of warming climate after 7-10 ka was followed by the

Holocene climatic optimum, which manifested in Australia as high lake levels (e.g. Bowler and Hamada, 1971; Gliganic et al., 2014; Wilkins et al., 2013), increased river discharge (Cohen and Nanson, 2007), reduced aeolian activity (as recorded in Blue Lake, in the Snowy Mountains; Stanley and De Deckker, 2002) and a maxima in rainforest and wet scelorphyll taxa suggesting a period of enhanced rainfall and temperatures 1-2° C higher than present (see Gouramanis et al., 2013 and; Kershaw, 1995 and references therein), between ~8.5-4 ka centered around 6 ka (Cohen and Nanson, 2007; Gouramanis et al., 2013). Thus, the most likely maximum age for the attainment of conditions suitable for organic soil building in the alpine area is considered to be 7-9 ka. This is likely to have had a considerable impact on the hydrology of alpine catchments as their capacity for storage and delayed release of rainfall and snowmelt increased with the development of organic soils and peatlands.

6.1.2 Evidence for late Holocene onset of soil development

The 2500 y cal. BP basal ages for the Guthega soils obtained by this study are concomitant with the Neoglacial, a period of glacial advances and enhanced geomorphic activity across the Southern Hemisphere (Porter, 2000). Evidence for altered climate conditions during this period is derived predominately from glacial advances, onset of peat development and palynological studies from southern South America and New Zealand (Kilian and Lamy, 2012; Markgraf et al., 1992; Porter, 2000; Shulmeister et al., 2004; Stansell et al., 2013; Wanner et al., 2008), in addition to Antarctic cooling at 5 ka (Hodell et al., 2001) and 2.3 ka (Masson-Delmotte et al., 2004). In the Southern Hemisphere, Neoglacial advances have widely been attributed to an increase in the strength/position of the southern westerly wind belt and an associated increase in precipitation (Lamy et al., 2001; Marx et al., 2011; Shulmeister et al., 2004; Strother et al., 2014). Glacial advances are typically taken as being centred around 4500 y cal. BP., however, a series of advances occur between 5400 and 2500 y cal. BP (Wanner et al., 2008).

The retreat of glaciers after 15 ka years in Australia means there is less obvious evidence of the Neoglacial in the Australian landscape compared with active glacial environments. However, the mid to late Holocene (after 6 ka) is associated with increased rainfall variability, including both droughts and pluvial episodes in association with the strengthening and increased frequency of the El Nino Southern Oscillation and the Indian Ocean Dipole (Gliganic et al., 2014; Gouramanis et al., 2013; Marx et al., 2011; Marx et al., 2009; Petherick et al., 2013). Evidence for cooler conditions is equivocal. Nevertheless, several distinct cold phases occurring throughout the mid to late Holocene in Australia are evident in $\delta^{18}O$ records from offshore records in the canyons of the Murray river at 4300, 2700 and 1400 years BP (Moros et al., 2009). These cold phases are generally replicated in deuterium derived records from the EPICA ice core (Masson-Delmotte et al., 2004). In addition there is evidence of cooler conditions in the Snowy Mountains at this time (e.g. Kemp and Hope, 2014; Martin, 1999), as discussed below.

In the Snowy Mountains, Costin (1972) found evidence of episodic geomorphic instability particularly at ~2000-4000 years BP. This included reactivation of solifluction terraces, rubble layers in fen peats surrounding Club Lake. The timing of this activity was established using some of the earliest radiocarbon dates undertaken in Australia, however, and occurred prior to advances in sample preparation to improve the removal of contamination.

More recent studies provide additional evidence of enhanced geomorphic activity and changing climate in the Australian Alps in the mid to late Holocene. Marx et al. (2011) found an increase in fluvial/colluvial sediment input to a peat mire in the headwaters of the Snowy River between 2000-4000 y cal. BP and a decrease in long-range dust input, interpreted to represent increased local geomorphic activity coinciding with wetter conditions in dust source areas (the lower Murray-Darling Basin).

The sedimentological pattern in lower elevation mires displays alternating peaty silt and humic clay bands between 10,000 and 3500 y cal. BP. At 3500 cal. BP, the peats are sometimes capped by a sand layer followed by continuous and rapid peat development to the present (Hope et al., 2009). Combined this indicates potentially variable geomorphic conditions before 3500 y cal. BP, potentially increased geomorphic activity at 3500 y cal. BP, followed by lower activity or wetter conditions after this date until the present. Further evidence is provided by higher elevation peat mires in the Snowy Mountains which at some sites began developing at c. 3000 y cal. BP, indicating the onset of favourable conditions for peat development at this time (Costin, 1972; Grover et al., 2012; Martin, 1999). The c.3000 y cal. BP signal in peat mires is potentially linked to periods of increased drought or fire (Hope et al., 2009). A possible decrease in temperature during this period is implied by palynological studies from high elevation sites. These show an increase in Poaceae pollen and a eucalypt minimum in the Diggers Creek peat bog from 1900-4000 y cal. BP (Martin 1999), a return to subalpine woodland or daisy rich grassland between 2700 and 900 y cal. BP at Micalong Swamp (Kemp and Hope, 2014) and an increase of Nothogagus at high elevation sites in the Victorian Alps (>900 m AHD) at approximately 2200 y cal. BP BP (McKenzie, 1997), which has been interpreted as indicating cooler and/or wetter conditions.

Consequently, there is evidence that supports the onset of alpine and subalpine soil development in the Snowy Mountains at 2500 y cal. BP. However, there is also evidence that soil formation, following the last Pleistocene glaciation, is likely to have begun from at least 7 ka. If the Guthega hillslope soils have developed from 2500 y cal. BP then it implies previous soil developed in the early to mid-Holocene has been eroded during the Neoglacial. Peat development continued through the Neoglacial period in the Snowy Mountains at high elevation sites, confirming significant biological activity. However, peat in the Snowy Mountains is confined to topographic depressions which are less likely to experience erosion. Overall, the age of soil development cannot be established unequivocally by this study; however, the minimum dates for the stable soil carbon fraction are consistent with the palaeoclimate evidence. The timing of the onset of soil formation and associated implications relating to the stability of the landscape remains important for understanding palaeoclimate conditions and geomorphic response in the Snowy Mountains.

6.2 Soil development rates

The current phase of soil development in the Snowy Mountains alpine zone has thus been proceeding for between 2.5 and 12 ka, producing between 30 and 60 cm of organic soil. The soil ages estimated by the radiocarbon ages and the geomorphic evidence provide two alternative methods for calculating the net rate of soil development (the net effect of soil production and removal) in the Snowy Mountains over the early to late Holocene. In both cases, estimates are likely the minimum rate at which soil is produced by weathering of bedrock, accumulation of mineral matter and dust accretion as soil mass is also influenced by erosion and soil creep. As discussed elsewhere in the manuscript these effects are likely to be minor however.

Assuming that soil development has been continuous since deglaciation i.e. since approximately 7-12 ka BP (method 1) produces in a mean estimated soil development (production minus removal) rate of approximately 0.05-0.09 mm/y (50 and 90 mm/ky) in the mid-slope and ridge-crest profiles and of 0.03–0.04 mm/y (30-40 mm/ky) in the toe-slope profile. Assuming an average bulk density of 0.7 g/cm^3 for the A horizon, this equates to a mean mass accumulation rate of 20–60 t/km^2/y.

Using the second approach (i.e. estimating soil development rates from the mass of soil accumulated between each radiocarbon dated interval using measured bulk densities of 0.6 g/cm^3 near the surface to 0.9 - 1.2 g/cm^3 at the base of the A horizon), soil development rates are estimated to have ranged between 0.08-0.26 mm/y or 40-220 t/km^2/y since the late Holocene. Thus the rate of soil development at Guthega estimated using the two methods described above ranges between a minimum of 0.05 and a maximum of 0.26 mm/y, with corresponding mass accumulation rates of 20 to 220 t/km^2/y.

Previously estimated rates for soil production undertaken in Australia using cosmogenic ^{10}Be nuclides ranged from 7 ± 0.2 mm/ky on conglomerate in semi-arid central Australia to 75 ±24 mm/ky on sandstone in Northern Territory tropics (original data from Heimsath et al., 2010, normalised to remove the effect of soil depth on soil development rates by Stockmann et al. 2007) The ^{10}Be method assumes steady state conditions where soil production equals denudation and so does not determine the separate production and erosion components. In contrast, production rates inferred from U-series isotopes are independent of this assumption and essentially measure the rate of downward migration of the soil/saprolite or saprolite/bedrock boundary. Soil production rates estimated using the U-series method at Frogs Hollow, NSW (930 AHD, 500 MAP), approximately 100 km northeast of the Snowy Mountains, were 10-24 mm/ky (for soils developed on Devonian granites) (Suresh et al., 2013). Assuming a bulk density of 1.6 g/cm^3, the combined range of reported production rates equate to a MAR for Australian soils of 11-120 t/km^2/y. Thus the range soil development (production minus erosion) rates in the Snowy Mountains reported in this study is towards the upper end or greater than previously reported soil production rates for Australian. This is despite the Snowy Mountains being traditionally viewed as a low chemical weathering environment.

The soil development rates estimated for Guthega hillslope (20-220 t/km^2/y) are broadly similar to those reported for Holocene aged soils on soil mantled alpine slopes elsewhere. Soil production rates estimated by the ^{10}Be method on the low relief plateau of the San Gabriel Mountains ranged from 38-196 t/km^2/y (Dixon et al., 2012). Dixon et al. (2014)considered these plateau sites to be relic landscapes unadjusted to tectonic uplift, making them somewhat comparable to the Snowy Mountains. Uplift rates in the San Gabriel Mountains are, however, rapid (i.e., 0.5-5 m/Ma), displaying a strong west-east gradient (see Lifton and Chase, 1992 and references therein), while MAP is four times lower than the Snowy Mountains, i.e. 500 mm/y.

The rates from both Guthega and the San Gabriel Mountains are also broadly comparable to those from the actively uplifting and relatively humid (MAP: 1137 mm) Rhone Valley (60-270 t/km^2/y) (Norton and von Blanckenburg, 2010). Soils of comparable age (1.5-12 ka) to those at Guthega have been shown to develop similar rates at various sites in the European Alps (32-346 /km^2/y) and on the Wind River Range, Wyoming (43-183 t/km^2/y) (Egli et al., 2014). Some of the highest rates of soil development have been measured in alpine areas experiencing both rapid uplift and very high annual precipitation. In the New Zealand Southern Alps (MAP 10 000 mm), for example, soil production rates reach 2.5 mm/y (Larsen et al., 2014), one to two orders of magnitude faster than the rates inferred for the Snowy Mountains.

The maintenance of significant soil mantles on alpine hillslopes, particular in wetter regions, has been attributed to the modulating influence of precipitation on the rate of soil production (Larsen et al., 2014). High moisture availability facilitates chemical weathering by supporting high vegetation productivity and consequently the production of chelating ligands, organic acids and increased subsurface CO_2 (Larsen et al., 2014; Riebe et al., 2004). By lowering the effective activation energy of weathering reactions, these biological processes may also lessen the otherwise inhibitory effect of low alpine temperatures on chemical weathering rates (Riebe et al., 2004).

In the Snowy Mountains, the high organic content of the chernic tenosols suggest an additional, more direct effect of vegetation in maintaining hillslope soil mantles. At 14-27 % the organic content of these soils greatly exceeds typical values for soils within Australia, where organic matter normally constitutes <5 % of the total soil weight (McKenzie et al., 2004). This difference is likely explained by high vegetation productivity in the Snowy Mountains as well as inhibition of organic decomposition by low temperatures, low pH and frequent saturation of the soil profile (McKenzie et al., 2004). Consequently, the SOM component alone directly equates to between 2 and 20 t/km^2/y of soil development in the Snowy Mountains. This is higher than estimates of SOM input for lowland Australian soils which are <1-6 t/km^2/y, (assuming a soil SOM content of 2-5 % and using the production rates of Heimsath et al., 2010; Stockmann et al., 2014)) Thus, organic input contributes significantly to the high soil production rates in the Snowy Mountains.

Removing the SOM component from the total soil development rate results in an estimated soil development rate for the Snowy Mountains of between 40-205 t/km^2/y. By comparison, estimates for lowland Australian soils, based on the data of Heimsath et al. (2010) are 11-113 t/km^2/y . Thus, at the upper end of estimates, soil production in the Snowy Mountains remains higher than other Australian rates implying the SOM contribution alone does not account for the high rates of Snowy Mountains soil production. Therefore rates of soil development in the Snowy Mountains may also be attributed to high weathering rates (i.e. conversion of bedrock to soil) and/or significant rates of dust accretion. High weathering rates may also be partly attributed to the high SOM content as organic acids contribute to mineral weathering (Larsen et al., 2014; Riebe et al., 2004) and high precipitation, which in the Snowy Mountains is 1/3 to four times higher than in other Australian sites where soil production has been examined).

6.3 Erosion rates

The $^{210}Pb_{ex}$ inventories from Guthega hillslope imply a hillslope erosion rate of 60 t/km^2/y with a range of 10-90 t/km^2/y over the past c.100 years. This compares to catchment-wide sediment yields of 13-23 t/km^2/y at Guthega and Club Lake. Sediment yields in this context represent minimum erosion estimates due to catchment storage and potential sediment loss from both lakes, Guthega Reservoir in particular. Statistical power analysis of the ^{137}Cs inventories demonstrated that the minimum erosion rate which could be detected given the sampling strategy was 70 t/km^2/y. No net soil erosion (or gain) was detectable using ^{137}Cs, implying erosion rates at Guthega hillslope are no higher than 70 t/km^2/y, which accords with upper range of modelled $^{210}Pb_{ex}$ erosion estimates. Combined these results indicate erosion rates at Guthega are likely to approximate those calculated using $^{210}Pb_{ex}$, i.e. ~60 t/km^2/y but may have be considerable lower, particularly during the past 60 years.

Snowy Mountains erosions rates (~60 t/km^2/y) are somewhat lower than the mean erosion rates (90 t/km^2/y) estimated for uncultivated Australian sites using the ^{137}Cs method (Chappell et al., 2011b; Loughran et al., 2004). By comparison, erosion rates estimated for the Southern Tablelands of Australia from sedimentation in farm dams range from 4 t/km^2/y for undisturbed sites to 90 t/km^2/y for degraded pasture (Neil, 1991). Assuming that the erosion rates estimated from the

$^{210}Pb_{ex}$ and sediment yield data are representative of the mean erosion rate for the 2500 km^2 subalpine/alpine zone as a whole, the total sediment yield from the Australian Alps over the past 9 ka approximates 540 x 10^6 t (although periods of significantly increased erosion, e.g. the Neoglaical, could result in considerably higher overall yields).

The $^{210}Pb_{ex}$ inventories and the lake sedimentation rates both incorporate the grazing era (mid 1800s-1940s CE), when high stocking rates for sheep and cattle lead to significant erosion in areas of the Snowy Mountains (Costin, 1954). Plot studies of soil erosion undertaken in the 1950s CE, immediately following the exclusion of grazing from the alpine area of the Snowy Mountains, found that erosion averaged around 460 ±220 (SE) t/km^2/y in areas where the ground cover had been disturbed by stock but was undetectable in areas where the vegetation cover remained intact (Costin et al., 1960) (Fig. 4).

In the Guthega catchment, soil surveys determined that an area of 30 km^{2,} out of the 90 km^2 catchment was slightly to severely eroded (Bryant, 1971). Notably, the hill slope studied here (Guthega hillslope) was not deemed to be significantly affected. Soil surveys reported the depth of erosion depth in addition to its spatial extent. Combined, these indicate a sediment yield of 180 t/km^2/y from affected areas. This equates to an increase in the background erosion rate (i.e. in addition to natural erosion from other parts of the catchment) of 60 t/km^2/y for the Guthega catchment (Fig. 5). Independent stream sediment gauging carried out at Guthega during the same period recorded a suspended sediment yield 18 t/km^2/y (Brown and Milner, 1988) (Fig. 5). The difference between these two estimates (18 versus 60 t/km^2/y) implies either that substantial quantities of sediment are stored within the catchment, or transported as bedload, or a combination of the two.

The $^{210}Pb_{ex}$ erosion rates presented by the current study are substantially lower than those measured during the grazing period. However, as approximately one half of the last 100 years, the time integrated by $^{210}Pb_{ex}$ erosion modelling, included grazing; the influence of grazing cannot be ruled out. While the study site was apparently not significantly affected by grazing, as indicated by observation based surveys (Bryant, 1971), it is possible that some grazing induced erosion did occur, albeit at a level no longer detectable by observation. If so, the erosion rates calculated here would exceed both the longer term late Holocene and post grazing rates, i.e. that the presented results overestimate typical Holocene soil loss rates. The fact that the Cootapatamba hillslope site recorded minimal to no net soil loss over the past 100 years does, however, imply that the effect of grazing must have been minimal on some sites, i.e. vegetated slopes as suggested by Costin et al. (1960).

Overall the magnitude of the increase in the rate of erosion initiated by the grazing era reinforces the importance of the protective vegetation cover in maintaining the hillslope soil mantle. It appears likely that high vegetation productivity in the Snowy Mountains facilitates the persistence of hillslope soils not only by contributing to rapid soil production but also through the protective effects of dense surface cover and subsurface rootmass in slowing erosion. This implies that the alpine region is sensitive to changes in climate or land conditions that can alter vegetation cover and result in increased net sediment erosion. Therefore, if cooling during the neoglacial was sufficient to reduce vegetation cover, even temporally (for example, vegetation composition is known to have changed during this period; Martin, 1999)then enhanced erosion (as previously discussed) seems likely.

7. Soil development, erosion and sedimentation – Summary and implications

The results of this study imply that the Snowy Mountains experience both rapid soil development rates and slow erosion rates by comparison to lowland sites. Maximum and minimum net soil development rates estimated by this study (20-220 $t/km^2/y$) exceed the maximum and minimum estimates of the net soil loss which has occurred over the past 100 years (10-90 $t/km^2/y$) (Fig. 5). This is consistent with the occurrence of widespread shallow alpine soils in the Snowy Mountains.

The $^{210}Pb_{ex}$ erosion rates measured by this study are, however, substantially lower than the rates of soil loss recorded during the grazing era (mid 1800s-1949 CE), during which the opening up of the protective vegetation cover lead to soil losses of 460 ± 220 t/km^2/y (Costin et al., 1960). This dichotomy demonstrates the importance of the dense vegetation cover to the persistence of the soil mantle on alpine hillslopes. Thus, currently, the rate of soil erosion in the Snowy Mountains appears to be outpaced by the processes of soil production. The minimum estimated date for the onset of the current soil production phase (2500 y cal. BP compared with the maximum estimated age of 7-12 ka), however, would imply that soil production has exceeded erosion in only the late Holocene, and could imply that significant soil erosion occurred on hill slopes during the neo-glacial period.

Under either scenario, the development rates estimated for the highly organic soils of the Snowy Mountains alpine area are high relative to those measured for lowland Australian soils (Heimsath et al., 2010; Stockmann et al., 2014). The Snowy Mountains rates are, however, comparable to production rates estimated for soil mantled hillslopes in several other alpine locations worldwide (Dixon et al., 2012; Egli et al., 2014). Rapid soil development in alpine areas has been attributed, previously, to the ready supply of fresh material in rapidly uplifting (eroding) areas and to the vegetation mediated enhancement of chemical weathering in high rainfall mountains (Larsen et al., 2014; Riebe et al., 2004). For tectonically stable, moderate rainfall alpine areas such, as the Snowy Mountains, it is suggested that high productivity of alpine vegetation provides significant material inputs and inhibits soil removal and may facilitate both rapid soil development and low sediment transport rates.

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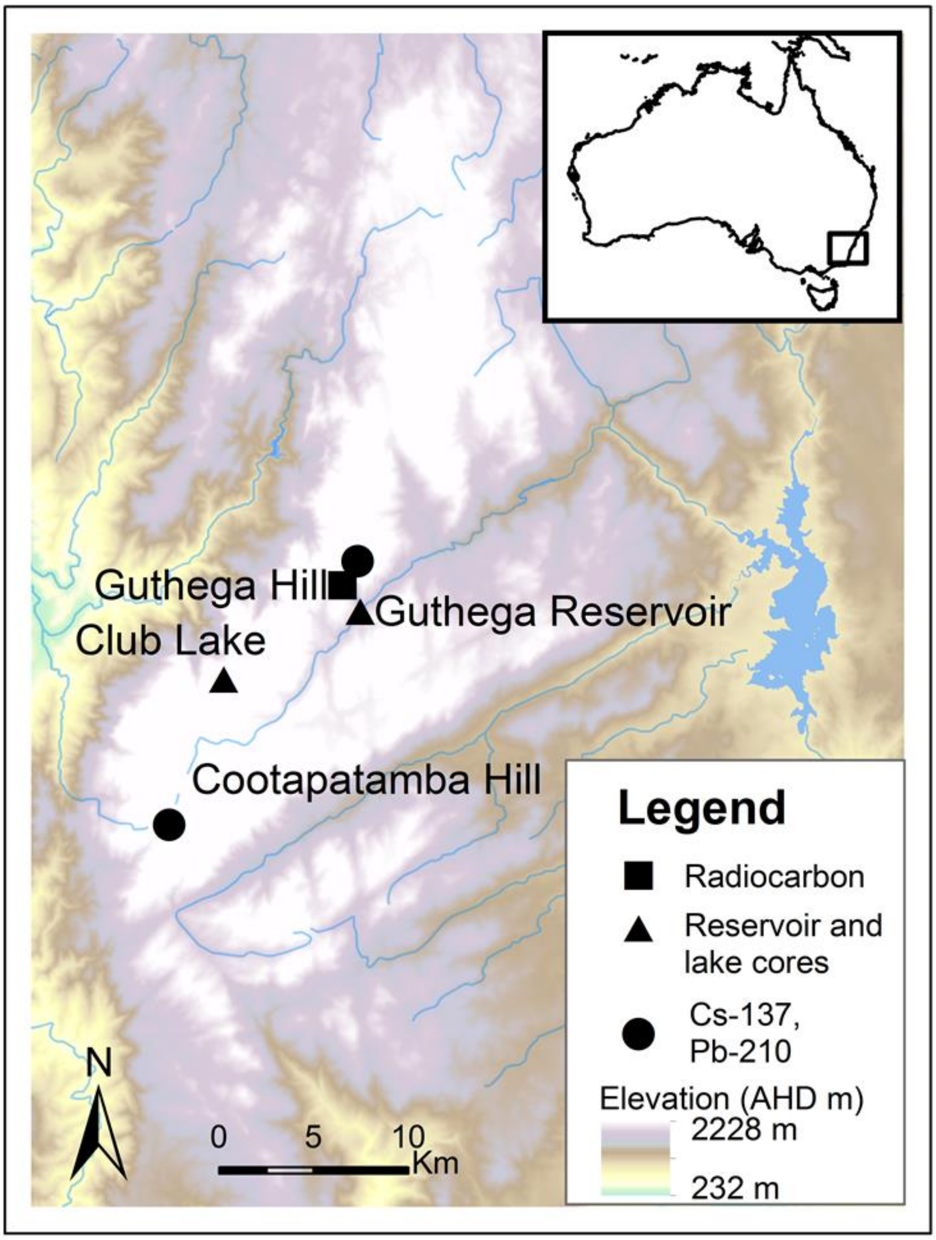

Fig. 1. Location of soil and sediment sampling sites.

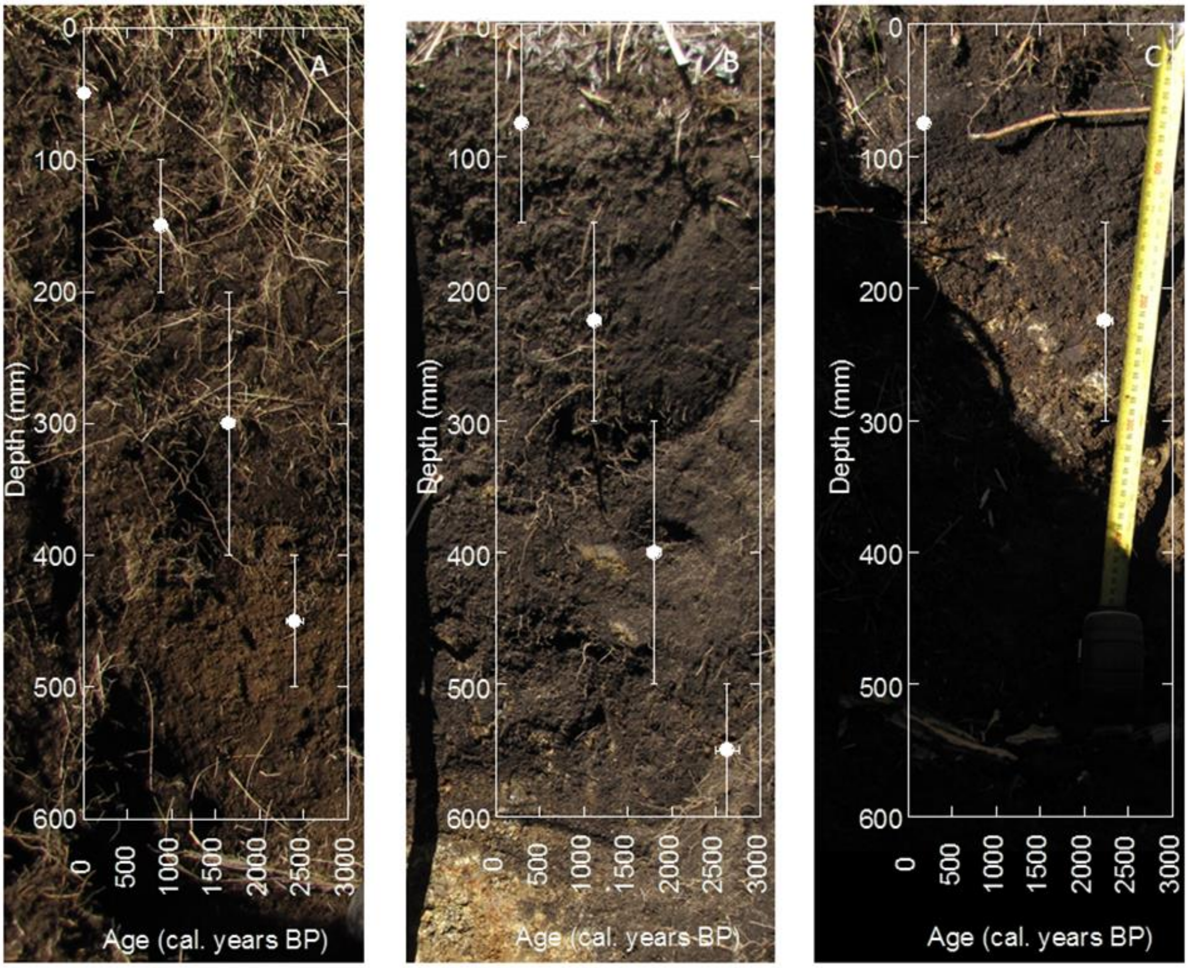

Fig. 2. Radiocarbon soil ages vs depth for soil pits at Guthega Hill: A) ridge-crest, B) mid-slope pit, C) toe slope.

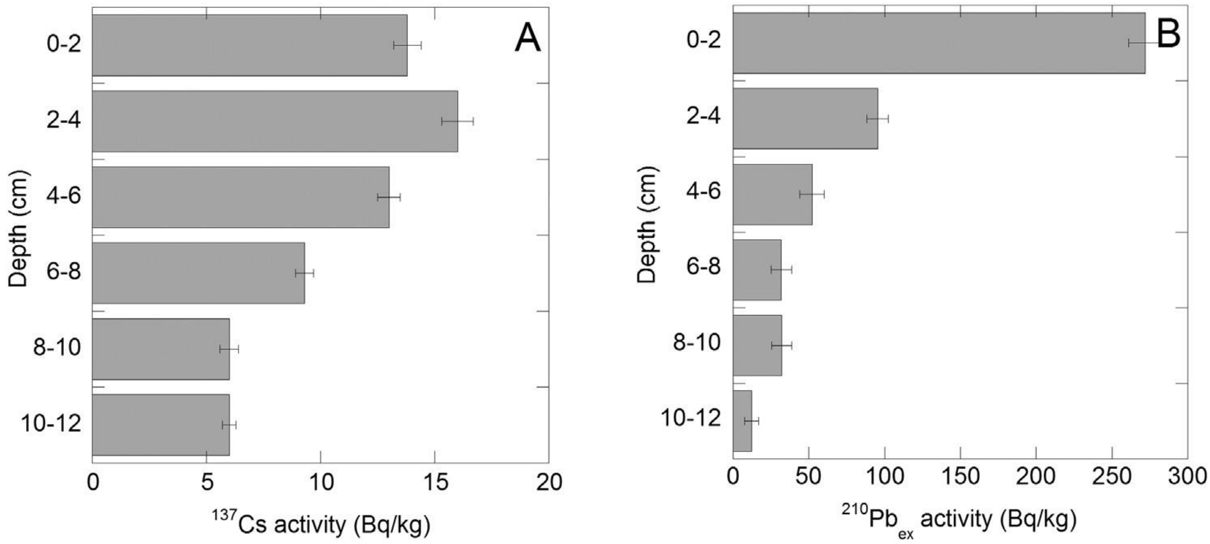

Fig. 3. Vertical distribution of (A)137Cs and (B) 210Pbex in Snowy Mountains soil. Plots show mean activity of 3 cores from Guthega reference site. Bars show counting error.

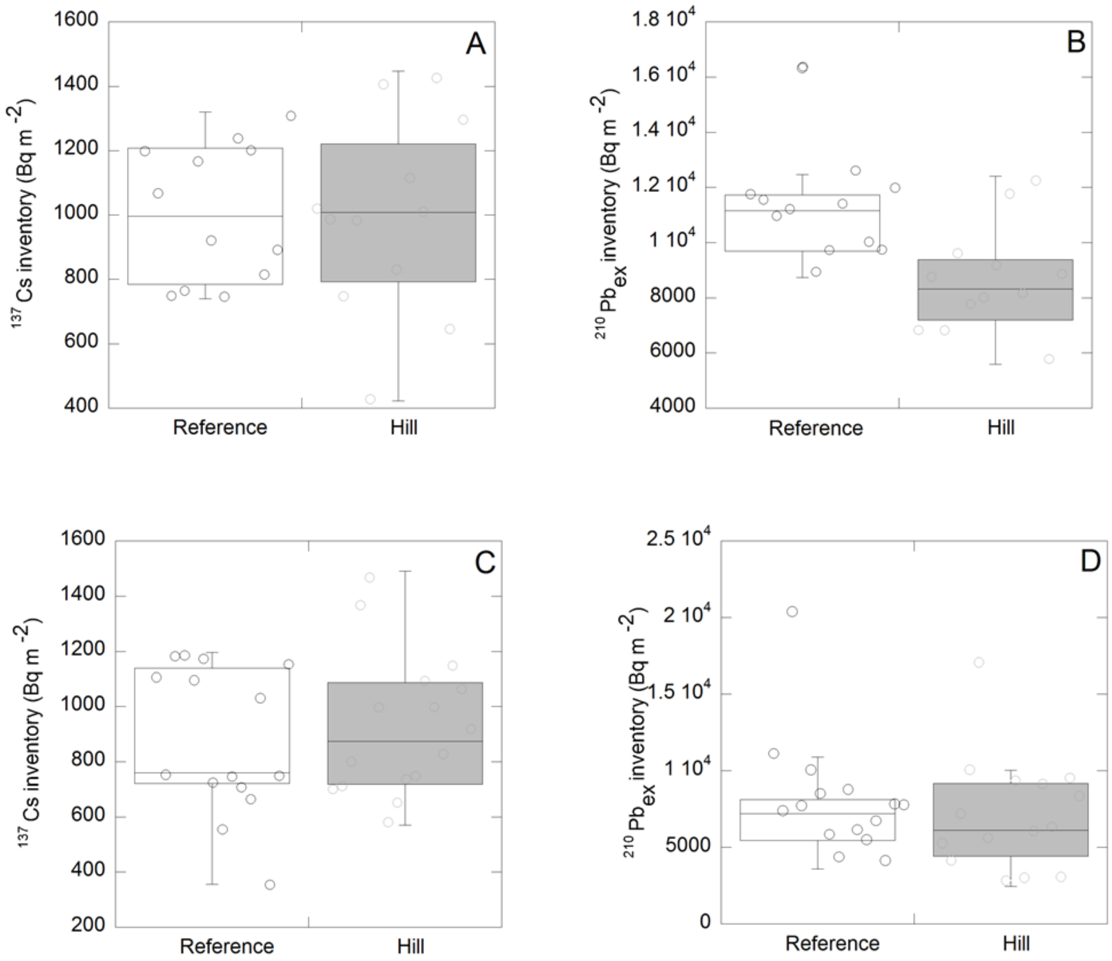

Fig. 4. Radionuclide inventories for reference and hillslope at Guthega A) 137Cs, B) 210Pbex and Cootapatamba C) 137Cs and D) 210Pbex.

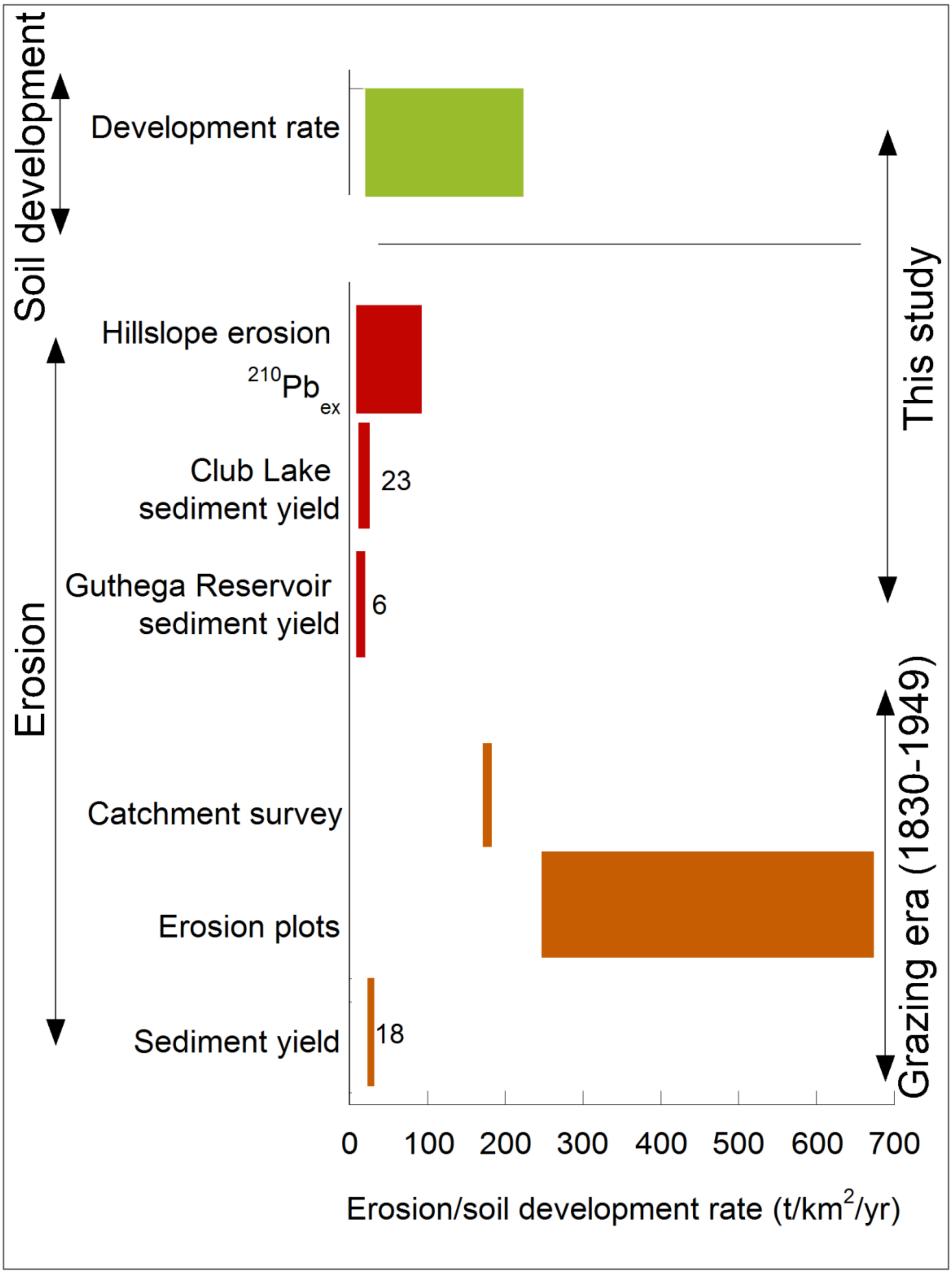

Fig. 5. Soil development and erosion rates estimated in this study comparedwith previous studies undertaken during the era of grazing induced erosion (catchment survey data from Bryant, 1971, erosion plot data from Costin et al., 1960, stream gauging data from Brown and Milner, 1988).

Table 1.

Model parameters describing radioactivity depth distribution for the profile distribution model (the equations for these parameters are provided in Walling et al. (2007)).

Site	Guthega Hill		Cootapatamba Hill	
Method	^{137}Cs	$^{210}Pb_{ex}$	^{137}Cs	$^{210}Pb_{ex}$
Relaxation depth (profile shape factor)a	4	4	4	4
Diffusion parameter (kg2m4yr1)b	36.96	2.69	17.2	0.83
Migration rate (kgm2yr1)c	0.62	0	0.42	0

a Describes the rate of decrease in radionuclide inventory with depth.

b Describes the penetration (producing dispersion) of the radionuclide into the soil profile with time.

c Describes the downward migration rate (downward shifting) of the radionuclide in the soil profile.

Table 2. Radiocarbon ages for hillslope soils

Lab code	Profile	Depth (mm)	Conventional Radiocarbon Age	Error[a]	Age years BP (calibrated)	Error[c]
Wk-39223	ridge-crest	0-100	104.5	± 0.3	7.66[b]	± 0.18
Wk-39224	ridge-crest	100-200	918	± 30	880	± 65
Wk-39225	ridge-crest	200-400	1790	± 25	1650	± 70
Wk-39226	ridge-crest	400-500	2392	± 26	2400	± 90
Wk-39228	mid-slope	0-150	268	± 25	295	± 25
Wk-39229	mid-slope	150-300	1234.00	± 25	1120	± 70
Wk-39230	mid-slope	300-500	1905.00	± 25	1800	± 80
Wk-39231	mid-slope	500-600	2584.00	± 29	2625	± 135
Wk-39233	toe-slope	0-150	222.00	± 25	185	± 45
Wk-39234	toe-slope	150-300	2251.00	± 25	2235	± 85

[a] 1 sigma standard deviation due to counting statistics multiplied by experimentally determined laboratory error multiplier

[b] This sample sample contained post-bomb levels of ^{14}C ($F^{14}C$ = 104.5%) and was calibrated using CALIBomb and the SHZ1-2 curve extension and the shcal 13 calibration dataset.

[c] 2 sigma standard deviation

Table 3. Descriptive statistics for radionuclide inventories at Guthega and Cootapatamba

Site	Method	Site type	Range	Mean	SE	95% confidence limit of the mean		Median	1/2 IQR	95% confidence limit of the median	
						Lower	Upper			Lower	Upper
Guthega	^{137}Cs	Reference[#]	740-1320	1008	63	1132	883	996	204	759	1209
		Hill slope	422	1447	88	1001	827	1009	180	750	1209
	$^{210}Pb_{xs}$	Reference	8734-16314	11199	564	10093	12305	11151	947	9545	11840
		Hill slope	5591-12400	8579	609	7385	9774	8322	918	6691	9618
Cootapatamba	^{137}Cs	Reference	356-1196	886	70	749	1023	759	226	713	1164
		Hill slope	570-1491	928	66	798	1727	874	177	713	1184
	$^{210}Pb_{xs}$	Reference	3597-20409	7692	1042	5649	9734	7185	1400	5308	8262
		Hill slope	2455-17385	6967	1002	1965	5002	8932	2378	3831	9297

[#]Reference site refers to the non-eroding site with which the hill slope sites are compared.

Table 4. Statisical tests for differences in distribution, mean and median between reference and hillslope radionuclide inventories for ^{137}Cs and $^{210}Pb_{ex}$

	Null Hypothesis	Test	^{137}Cs		$^{210}Pb_{ex}$	
			Signficance (p)	Hillslope and reference inventories are signficantly different (Y/N)	Signficance (p)	Hillslope and reference inventories are signficantly different (Y/N)
Guthega	A#	t-test	0.953	N	0.004	Y
	B*	ISMT[2]	1	N	0.03	Y
	C[1]	ISKST[3]	0.996	N	0.01	Y
Cootapatamba	A#	t-test	0.664	N	0.62	N
	B*	ISMT[2]	0.862	N	0.953	N
	C[1]	ISKST[3]	0.876	N	0.925	N

A#= the mean inventory at the hillslope is the same as would be expected due to fallout and natural variability alone

B*= the median inventory at the hillslope is the same as would be expected due to fallout and natural variability alone

C[1]= the distribution of the hillslope inventory is the same as would be expected due to natural variability alone

ISMT[2] = independent samples median test

ISKST[3] = independent samples Kolmogorov Smirmov test