

Meteoric beryllium-10 fluxes from soil inventory measurements in the East River watershed, Colorado, USA

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Abstract. Meteoric beryllium-10 ($^{10}\text{Be}_{\text{met}}$) has a wide range of applications as a geochronometer and tracer of geological processes. $^{10}\text{Be}_{\text{met}}$ is produced in the atmosphere by cosmic rays and delivered to Earth's surface primarily via precipitation. $^{10}\text{Be}_{\text{met}}$ is particularly suitable for quantifying surface process rates where use of *in situ*-produced ^{10}Be is challenging, such as landscapes with quartz-poor bedrock. However, using $^{10}\text{Be}_{\text{met}}$ for dating and quantifying surface process rates requires constraining depositional fluxes across space and time. Although empirical and physical models for predicting fluxes exist, the predictions can deviate substantially from measured values. Here we quantify $^{10}\text{Be}_{\text{met}}$ flux in the East River watershed in Colorado, USA where precipitation is dominated by snowfall. We measured the $^{10}\text{Be}_{\text{met}}$ inventory in soils on five glacial moraines 13–18 ka in age that span 700 m of elevation and calculated $^{10}\text{Be}_{\text{met}}$ fluxes by dividing each inventory by moraine age. Inheritance-corrected fluxes range from 4.31×10^5 – 1.80×10^6 atoms $\text{cm}^{-2} \text{yr}^{-1}$ and fluxes corrected for erosion are higher by up to a factor of two. The predicted fluxes are well correlated with elevation, mean annual precipitation, mean snow depth, and snow water equivalent ($R^2 = 0.82$ to 0.98), but the relationships degrade when fluxes are corrected for erosion ($R^2 = -0.42$ – 0.63). Regression models based on elevation, precipitation, snow depth and snow water equivalent predict watershed-averaged fluxes of 0.66×10^6 – 2.31×10^6 atoms $\text{cm}^{-2} \text{yr}^{-1}$. Predicted fluxes from a published empirical model that estimates fluxes as a function of precipitation were within a factor of 1.8–3.8 of measured values at each site. Fluxes predicted by physically-based general circulation models (GCM) are generally within a factor of one to three of our estimated watershed-averaged values, but the GCM predictions are too coarse to capture the intra-watershed spatial variability in fluxes. Our results highlight both the importance of factors that drive variability in $^{10}\text{Be}_{\text{met}}$ delivery to soils and how local calibration can improve estimates of $^{10}\text{Be}_{\text{met}}$ flux in mountain watersheds.

1 Introduction

The cosmogenic nuclide beryllium-10 (^{10}Be) is used widely for geochronology and to quantify rates of Earth surface processes (Bierman and Nichols, 2004; Willenbring and von Blanckenburg, 2010; Granger et al., 2013). ^{10}Be forms during high energy collisions between cosmic particles and target nuclei (Lal and Peters, 1967). ^{10}Be produced in Earth's atmosphere (typically

from N and O) is known as atmospheric, garden variety, or meteoric ^{10}Be ($^{10}\text{Be}_{\text{met}}$), whereas *in situ*-produced ^{10}Be ($^{10}\text{Be}_{\text{in situ}}$) is formed in mineral lattices at and near Earth's surface. Production rates of $^{10}\text{Be}_{\text{in situ}}$ are on the order of a few to tens of atoms $\text{g}^{-1} \text{yr}^{-1}$ in quartz (Gosse and Phillips, 2001; Borchers et al., 2016). In contrast, $^{10}\text{Be}_{\text{met}}$ is generated at orders of magnitude higher rates, with global-averaged flux values in excess of one million atoms $\text{cm}^{-2} \text{yr}^{-1}$ (Monaghan, et al., 1986; McHargue and Damon, 1991). $^{10}\text{Be}_{\text{met}}$ is circulated throughout the atmosphere before being removed through precipitation or dry fallout (Field 2006; Heikillä et al., 2008). Once delivered to Earth's surface, $^{10}\text{Be}_{\text{met}}$ binds with minerals within soil, and is preferentially adsorbed to the surfaces of fine-grained particles and secondary minerals (e.g., Graly et al., 2010). $^{10}\text{Be}_{\text{met}}$ is partitioned with the solid phase for approximately neutral pH values (You et al., 1989) and is generally thought to be retained in soils where pH is $> \sim 4$ (Graly et al., 2010). However, $^{10}\text{Be}_{\text{met}}$ may be lost from upper soil horizons where low pH leads to leaching (Dixon et al., 2018), where clay minerals or humic material with adsorbed Be are translocated to greater depth (Pavich et al., 1984; Jagercikova et al., 2015), and where the surface of the soil has been eroded (Harden et al., 2002). Hence soil $^{10}\text{Be}_{\text{met}}$ inventories depend on a variety of factors, including the local delivery rate, landform age, degree of chemical weathering, and erosion rate (e.g., Graly et al., 2010; Willenbring and von Blanckenburg, 2010).

The high production rates, which allow $^{10}\text{Be}_{\text{met}}/^9\text{Be}$ ratios to be readily measured via accelerator mass spectrometry (AMS), the wide range of soils and regolith where $^{10}\text{Be}_{\text{met}}$ is retained, and – unlike $^{10}\text{Be}_{\text{in situ}}$ – the ability to be measured in non-quartz-bearing materials has allowed $^{10}\text{Be}_{\text{met}}$ to be used widely as a geochronometer and tracer of Earth surface processes (e.g., Brown et al., 1987a; Graly et al., 2010; Willenbring and von Blanckenburg, 2010). For example, $^{10}\text{Be}_{\text{met}}$ has been used to date terraces and moraines (Tsai et al., 2018; Elgi et al., 2010; Dünforth et al., 2012), determine rock-to-soil production rates (Monaghan et al., 1992; McKean et al., 1993; Jungers et al., 2009), quantify soil residence times (Monaghan et al., 1983; Bacon et al., 2012; Foster et al., 2015), determine earthflow velocities (Macky et al., 2009), trace sediment through hillslopes and fluvial systems (Brown et al., 1988; Ruesser and Bierman, 2010; West et al., 2014; Campforts et al., 2016), and, when coupled with ^9Be measurements, to quantify watershed-scale denudation rates and chemical weathering (von Blanckenburg et al., 2012; 2014; Dannhaus et al., 2018; Portenga et al., 2019; Deng et al., 2020a; 2021). However, a requirement for using $^{10}\text{Be}_{\text{met}}$ for dating and quantifying surface process rates is that the deposition rate or flux of $^{10}\text{Be}_{\text{met}}$ (F_{met} , $\text{atom cm}^{-2} \text{yr}^{-1}$) to the study area is known.

Numerous efforts have been made to measure or predict F_{met} at local to global spatial scales and over annual to millennial timescales (Willenbring and von Blanckenburg, 2010; Graly et al., 2011). Direct measurement of $^{10}\text{Be}_{\text{met}}$ concentrations in precipitation has been used to quantify $^{10}\text{Be}_{\text{met}}$ flux at point scales and short timescales to develop empirical relations that predict F_{met} as a function of precipitation rate or latitude (Brown et al., 1989; Monaghan et al., 1986; Raisbeck et al., 1981; Graham et al., 2003; Graly et al., 2011; Willenbring and von Blanckenburg 2010). Global-scale values of F_{met} are predicted with nuclide production functions (e.g., Lal and Peters, 1967; Masarik and Beer, 1999) that are coupled with general circulation models (GCM) to simulate spatial and temporal patterns of F_{met} as a function of geomagnetic field strength, solar activity,

65 atmospheric circulation, and precipitation (Field et al., 2006; Heikillä et al., 2008; Heikillä and von Blanckenberg, 2015; Panovska et al., 2023; Zheng et al., 2021; Zheng et al., 2024). Millennial-averaged values of F_{met} can be inferred from soils by first measuring the $^{10}\text{Be}_{met}$ inventory (I_{met} , atom cm^{-2}), which is done by multiplying the inheritance-corrected $^{10}\text{Be}_{met}$ concentration (N_{corr} , atom g^{-1}), by soil bulk density (ρ , g cm^{-3}), the thickness of the sampling interval (z , cm), and the volumetric fraction of fine soil V_f , and summing the values for all (n) individual sample depth increments (i) (e.g., Deng et al., 2021):

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$$I_{met} = \sum_{i=1}^n (N_{corr} \cdot \rho \cdot z \cdot V_f)_i \quad (1)$$

Provided the landform age is much less than the mean life of ^{10}Be , the I_{met} value measured on an independently dated landform is then divided by age (t , yr),

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$$F_{met} = \frac{I_{met}}{t}, \quad (2)$$

to determine the flux (e.g., Graly et al., 2010; Reusser et al., 2010; Ouimet et al., 2015; Clow et al., 2020).

80 Using soil inventories to quantify F_{met} assumes that $^{10}\text{Be}_{met}$ accumulation begins at the time a landform is initially exposed to wet and dry fallout. However, the ages of most landforms used to constrain flux are late Quaternary in age, or 1–2 orders of magnitude lower than the ^{10}Be half-life (1.387 Ma; Chmeleff, et al., 2010; Korschinek et al., 2010). Hence most soil inventories contain inherited $^{10}\text{Be}_{met}$ atoms that were adsorbed to sediment grains prior to their incorporation within terraces and moraines (Graly et al., 2011). The inherited component can be identified by the presence of a near constant $^{10}\text{Be}_{met}$ concentration in the
85 lower part of a measured depth profile and, once identified, can be subtracted from measurements at shallower depths to determine inheritance-corrected inventories (e.g., Reusser et al., 2010).

An additional assumption required to calculate F_{met} from soil inventories is that all $^{10}\text{Be}_{met}$ delivered to the soil since landform initiation has been retained (Willenbring and von Blanckenburg 2010). Potential for $^{10}\text{Be}_{met}$ loss due to chemical leaching can
90 be assessed by measuring soil pH values. Potential loss of $^{10}\text{Be}_{met}$ due to physical erosion can be minimized by selection of stable, flat surfaces for sampling. However, if the surface erosion rate is independently known, the flux can be determined for eroding landforms, provided the $^{10}\text{Be}_{met}$ concentration in eroded material is also known (Brown, 1987b; Clow et al., 2020).

Soils contain $^{10}\text{Be}_{met}$ scavenged from the atmosphere by aerosols, but also from dust particles that are derived from aeolian
95 erosion and transported in the atmosphere before being deposited (Monaghan et al., 1986; Graly, 2011). The dust-derived or ‘recycled’ $^{10}\text{Be}_{met}$ can be as much as ~20% of the total flux (Monaghan et al., 1986; Graham et al., 2003) and must be accounted for before comparing measured fluxes against predictions from GCMs. However, the dust contribution to flux estimates does

not need to be independently known to quantify surface process rates, provided that time-averaged delivery of dust to soils is representative of the timescale relevant to the surface process rate under investigation.

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There is often disagreement between F_{met} values determined from soil inventories and global-scale physical models or empirical models that predict flux as a function of precipitation and latitude (Ouimet et al., 2015; Deng et al., 2020b; Krone et al., 2024). Substantial variability in F_{met} has been documented at spatial scales smaller than the grid cell size (often several hundred km) of many GCMs (Ouimet et al., 2015; Clow et al., 2020; Deng et al., 2020b). The discrepancies and spatial variability likely stem from temporal and spatial fluctuations in $^{10}\text{Be}_{met}$ production, atmospheric circulation, and precipitation (Monaghan et al., 1986; Graham 2003; Jordan et al., 2003; Deng et al., 2020a; Krone et al., 2024). Hence, it is often desirable to locally-calibrate F_{met} prior to using $^{10}\text{Be}_{met}$ to quantify surface process rates (e.g., Reusser et al., 2010; Ouimet et al., 2015). Local-scale calibration is particularly important in mountainous settings, where microclimate variability can cause $^{10}\text{Be}_{met}$ inventories and fluxes to vary over short distances (Ouimet et al., 2015). Here we quantify the $^{10}\text{Be}_{met}$ flux at sites throughout the East River watershed, near Crested Butte, Colorado, USA (Figure 1), a site of intensive research on the physical and biogeochemical drivers of watershed function (Hubbard et al., 2018). We measure $^{10}\text{Be}_{met}$ inventories on independently dated glacial moraines, assess local-scale influences on $^{10}\text{Be}_{met}$ flux, and compare $^{10}\text{Be}_{met}$ flux measurements against predictions from empirical and physical models to evaluate the spatial variability and predictability of F_{met} values in alpine environments.

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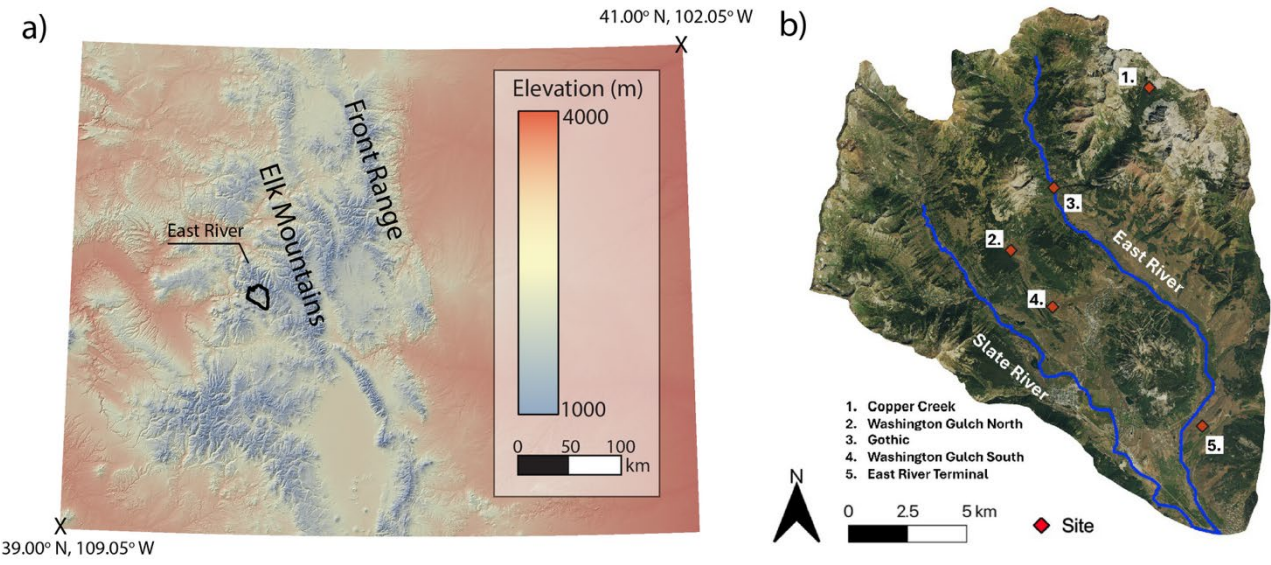


Figure 1: a) Digital elevation model (DEM) of Colorado, USA showing the location of East River watershed. b) Image of the East River watershed showing locations (red diamonds) where soil $^{10}\text{Be}_{met}$ inventories and fluxes were measured.

2 Study Site

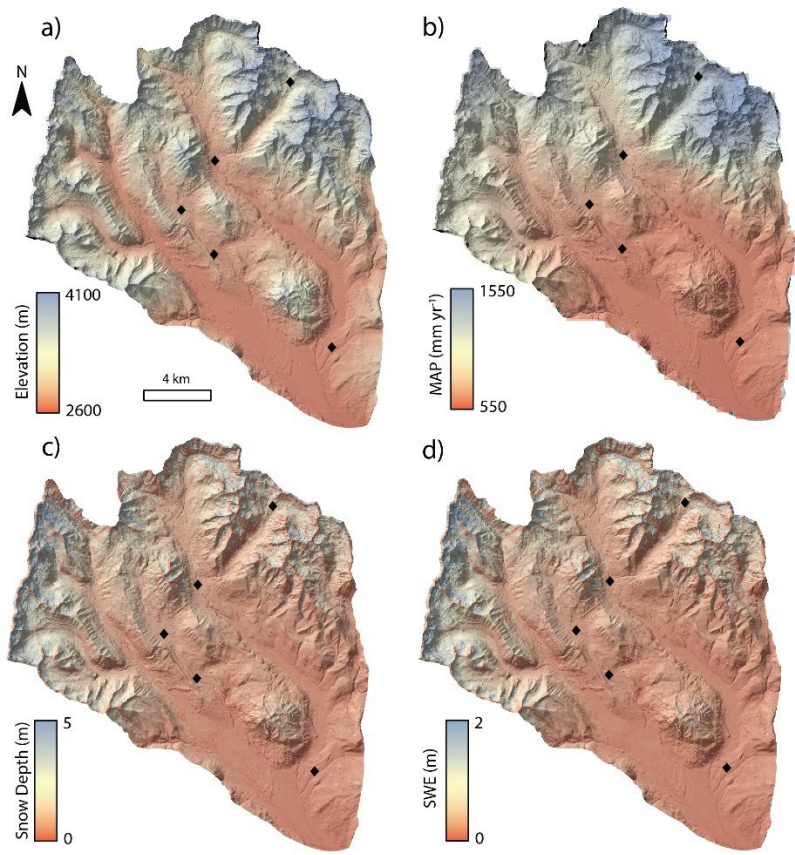


Figure 2: Maps of environmental variables within the East River watershed. (a) LiDAR-derived elevation data; (b) Mean annual precipitation (MAP) averaged from 2008–2019; (c) Mean snow depth, and (d) Mean snow water equivalent (SWE), derived from Airborne Snow Observatory data from 2018–2019. The black diamonds show each sample location.

The East River drains the Elk Range of the Rocky Mountains in Colorado, USA (~38.9°N, 106.9°W). The East River joins the Taylor River to form the Gunnison River, a major tributary to the Colorado River. The East River watershed and its tributaries, Washington Gulch and the Slate River, are a steep mountainous landscape, with elevations that range from 2600 m to 4100 m above sea-level (Figure 2a). The climate is characterized by long, cold winters and cool, short summers (Hubbard et al., 2018). The mean annual temperature is ~0 °C, with mean minimum and maximum temperatures of -9.2 and 9.8 °C, respectively (Hubbard et al., 2018). Most of the annual precipitation occurs as snowfall, which can exceed several meters per year (Prather et al., 2023) and varies with elevation (Figure 2). The continental subarctic climate in the East River watershed

supports montane, subalpine, and alpine ecosystems, which host aspen, meadow, mixed conifer, sagebrush, grass, sedge, and
135 forb species (Figure S1) (Harte and Shaw, 1995). Bedrock in the watershed is primarily Cretaceous-age Mancos Shale and
other sedimentary rocks that are intruded by Paleocene- to Miocene-age quartz monzonite porphyry and granodiorite rocks
(Gaskill et al., 1967, 1991). Pleistocene glaciers eroded high elevation cirques and deposited moraines and outwash in the main
East River valley (Gaskill et al., 1967, 1991). Exposure ages based on $^{10}\text{Be}_{in situ}$ concentrations measured in samples from
140 boulders indicate the terminal moraine at the end of the valley dates to 17.9 ka, whereas the youngest recessional moraine in a
high-elevation cirque within the Copper Creek watershed was exposed at 13.0 ka (Quirk et al., 2024).

3 Methods

3.1 Field and laboratory methods

Soil $^{10}\text{Be}_{met}$ inventories were determined for five glacial moraines with exposure ages that range from 17.9 to 13.0 ka (Figure
1; Table 1; Quirk et al., 2024). Soil pits were dug on the flattest part of each moraine crest to minimize the potential of post-
145 depositional erosion, with the exception of the Gothic moraine site, which was opportunistically located where the East River
has laterally eroded a cutbank into the moraine, permitting deeper sampling. The pits were dug until shovel refusal, to depths
between 60 and 145 cm. At the Washington Gulch North site, we sampled from a soil pit on a flat portion of the moraine, and
collected a sample on the same moraine that was exposed by a recent landslide ~375 m from the soil pit to assess inheritance
in a deeper sample than we could access by digging. Samples were collected in either 5 or 10 cm depth increments by pressing
150 a stainless-steel box, open at the top and front, into the face of the soil pit and scraping soil into the box. Samples were dried
and sieved to separate < 2 mm particles. We combined samples from adjacent depth intervals to generate composite samples
that encompassed depth intervals that range from 10 to 20 cm. Samples were combined by mixing splits of the < 2 mm size
fraction in proportion to the total mass of < 2 mm particles in each sampling interval. The combined < 2 mm fractions were
powdered using a tungsten carbide shatter box. We measured soil pH to assess $^{10}\text{Be}_{met}$ retention by first mixing ~5 g of
155 powdered < 2 mm grain-size material with 12 mL ultrapure water. Samples were stirred and left overnight prior to
measurements with a pH probe.

Table 1. Age and location of the calibration sites used to assess meteoric ^{10}Be flux.

Moraine	Age ka	Latitude	Longitude	Slope (°)	Curvature (m ⁻¹)
Copper Creek	13.0±0.1	39.001190	-106.940171	7.0	0.00
Washington Gulch North	17.9±0.2	38.930916	-107.013397	2.0	-0.03
Gothic moraine	14.0±1.0	38.957877	-106.990814	4.3	-0.04
Washington Gulch South	17.3±0.1	38.907394	-106.990058	0.5	-0.01

East River terminal	17.8±0.2	38.858632	-106.907514	4.0	-0.02
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160 See Quirk et al. (2024) for exposure age details. Slope (°) and curvature values were generated using 3-meter LiDAR DEM (Goulden et al., 2020). Curvature (m^{-1}) was derived by fitting a polynomial surface to a neighbourhood of 11 pixels using LSDtopotools (Mudd et al., 2014). Locations were determined using a Bad Elf Flex GNSS with sub-meter precision.

165 $^{10}\text{Be}_{\text{met}}$ chemistry was conducted at the University of Massachusetts Cosmogenic Nuclide Laboratory using the Stone (1998) fusion method. Approximately 0.5 g of the homogenized and pulverized < 2 mm soil particles was weighed into 30 mL Pt crucibles. We then added ~350 μg of ^9Be from a carrier with a $^{10}\text{Be}/^9\text{Be}$ ratio of $\sim 1.75 \times 10^{-15}$ to each sample. We also added 1000 μg of Al from a commercial carrier, which reportedly improves Be yields (Stone and Balco, 2004). Samples were dried overnight at 60 °C, then mixed with ~2.5 g of KHF_2 and ~0.5 g of NaSO_4 flux prior to heating with a gas torch to melt the

170 sample. Once the samples were fully melted, they were heated for an additional 1–2 minutes. Be was extracted by soaking the crucibles with the fusion cakes overnight in 200 mL of ultrapure water. The solution was reduced to a volume of ~30 mL by evaporation, then centrifuged. We decanted the supernatant and added ~10–15 mL of 49% HClO_4 to precipitate K as KClO_4 . The solution was centrifuged to remove KClO_4 and the supernatant was dried overnight in Teflon beakers following addition of 1 mL of 8M HNO_3 . The samples were redissolved in 25 mL ultrapure water and Be was precipitated as BeOH by adding

175 drops of ~10% NH_4OH until the solution pH was ~9. The BeOH was washed multiple times by repeatedly centrifuging, pouring off the supernatant, and remixing the BeOH with ultrapure water. The BeOH was dissolved with 2-3 drops 6M HCl , dried overnight in a quartz crucible, calcined by heating with a gas torch until BeO remained incandescent for 1 minute, and packed into cathodes with Nb powder. The $^{10}\text{Be}/^9\text{Be}$ ratios were analyzed by AMS at Lawrence Livermore National Laboratory (LLNL) against AMS Be standard 01-5-4 (LLNL standard 07KNSTD) (Nishiizumi et al., 2007). ^{10}Be concentrations in soil

180 samples were blank-corrected by subtracting $^{10}\text{Be}/^9\text{Be}$ ratios measured in full process blanks from ratios measured in samples, with uncertainties added in quadrature.

3.2 $^{10}\text{Be}_{\text{met}}$ inventory and flux analysis

We calculated $^{10}\text{Be}_{\text{met}}$ inventories following equation 1 using blank- and inheritance-corrected $^{10}\text{Be}_{\text{met}}$ concentrations, bulk density values measured for different soil horizons, and estimates of the < 2 mm clast volume. Bulk density was determined

185 by driving a 90.8 cm^3 steel cylinder into the soil at depths corresponding to different horizons and weighing the dry soil (Table S1). Bulk density was assigned to each $^{10}\text{Be}_{\text{met}}$ depth interval based on soil horizons identified in the field (Table S2). The volumetric fraction of < 2 mm soil particles was calculated for each depth interval following Deng et al. (2021) as: $V_f = (1 - V_{c1}) \times (1 - V_{c2})$, where V_{c1} and V_{c2} are volumes of two sizes of coarse clasts. V_{c1} was determined by sieving each sample to separate < 2 mm and > 2 mm clasts and the volume fraction of > 2 mm clasts was calculated assuming a density of 2.65 g cm^{-3} .

190 ³. V_{c2} was estimated by using field photos of each soil pit to digitize the area of large cobble- to boulder-sized clasts that were too large to be sampled by our 44.7 mm diameter steel ring. The volume fraction of large clasts was calculated by multiplying the area fraction by two-thirds (e.g., Deng et al., 2021).

At four of the five sites, the $^{10}\text{Be}_{\text{met}}$ concentration reached a near-constant value with depth (Figure 3). We assumed the concentration measured in the lowest depth interval was equal to the inherited $^{10}\text{Be}_{\text{met}}$ concentration and subtracted that value from concentrations measured in overlying samples (e.g., Ruesser et al., 2010) and added the 1σ AMS analytical uncertainties in quadrature. At the fifth site in the Copper Creek watershed, the $^{10}\text{Be}_{\text{met}}$ profile is well-mixed above an indurated stone line at the base of the soil (Figure 3a). Hence, we used the mean inheritance value of 7.58×10^7 atoms g^{-1} measured at the other four sites to correct for inheritance at the Copper Creek site. If the $^{10}\text{Be}_{\text{met}}$ concentration in the sample used for the inheritance correction was greater than the concentration of a sample higher in the profile (e.g., at the Gothic moraine site), we set the concentration for that depth interval to zero. We propagated uncertainty for individual depth increments by adding uncertainty associated with the inheritance-corrected $^{10}\text{Be}_{\text{met}}$ concentrations and an assumed 5% relative uncertainty in both ρ and V_f in quadrature. Uncertainty in I_{met} values was determined by adding the uncertainty for individual depth increments in quadrature. Analytical uncertainties were low (1.3-2.2%) and the I_{met} uncertainties primarily reflect the uncertainties in ρ , V_f , and in some cases, the inheritance-corrected $^{10}\text{Be}_{\text{met}}$ concentration. In some cases, uncertainties in the inheritance-corrected $^{10}\text{Be}_{\text{met}}$ concentrations were large relative to the $^{10}\text{Be}_{\text{met}}$ concentrations because the difference in concentrations between the two depth increments was small. Hence the Washington Gulch North and Gothic inventories have slightly higher uncertainties relative to the other sites.

Inheritance-corrected F_{met} values were calculated using equation 2, assuming no erosion. Uncertainty in F_{met} for each calibration site was calculated by adding uncertainty in I_{met} values and moraine age in quadrature. Moraine age uncertainty was based on the standard error of the mean of multiple exposure ages or, for the Gothic moraine, a 1 ka uncertainty associated with a $^{10}\text{Be}_{\text{in situ}}$ depth profile age (Quirk et al., 2024).

The $^{10}\text{Be}_{\text{in situ}}$ depth profile method used to date the Gothic moraine, which was also sampled for $^{10}\text{Be}_{\text{met}}$, yielded a median erosion rate of 0.023 mm yr^{-1} (Quirk et al., 2024). The source area for the Gothic moraine is primarily intrusive rock, hence the till matrix contained sufficient quartz for $^{10}\text{Be}_{\text{in situ}}$ analysis. However, at the other sites, the till matrix was dominated by quartz-poor Mancos shale and because of the dense shale-derived matrix, the pits could not be excavated deep enough to permit use of a $^{10}\text{Be}_{\text{in situ}}$ profile to assess erosion rates. We assessed the potential magnitude of the influence of erosion on fluxes by calculating an erosion-corrected flux ($F_{\text{cor}_\varepsilon}$) as:

$$F_{\text{cor}_\varepsilon} = F_{\text{met}} + (\varepsilon \cdot N_{\text{surface}}) + (I_{\text{met}} \cdot \lambda) \quad (3)$$

where ε is the erosion rate ($\text{g cm}^{-2} \text{ yr}^{-1}$), N_{surface} is the $^{10}\text{Be}_{\text{met}}$ concentration in the uppermost sample, and λ is the ^{10}Be decay constant ($4.998 \times 10^{-7} \text{ yr}^{-1}$). For low-gradient hillslopes like the moraines we sampled, erosion rates (ε) scale linearly with topographic curvature,

$$\varepsilon = \frac{dz}{dt} = D \nabla^2 z, \quad (4)$$

where dz/dt is the change in elevation (z) per unit time (m yr^{-1}), D is the topographic diffusion coefficient ($\text{m}^2 \text{ yr}^{-1}$), and ($\nabla^2 z$) is topographic curvature (m^{-1}) (e.g., Kwang et al., 2022). We calculated topographic curvature at each site using LSDTopoTools (Mudd et al., 2014). Based on the median erosion rate ($-2.3 \times 10^{-5} \text{ m yr}^{-1}$) and topographic curvature (-0.04 m^{-1} ; Table 1), we infer the value of D at the Gothic site is $5.8 \times 10^{-4} \text{ m}^2 \text{ yr}^{-1}$. Given the similar soils and climate, we assume D is comparable across the five moraines and use the diffusion coefficient for the Gothic site and topographic curvature values to estimate erosion rates at the four sites that lack direct measurements. We also modeled $F_{\text{cor}_\varepsilon}$ using ε values ranging from 0 to 0.04 mm yr^{-1} to encompass the range of possible median erosion rates (0.016 – 0.036 mm yr^{-1}) inferred from the $^{10}\text{Be}_{\text{in situ}}$ depth profile at the Gothic moraine site (Quirk et al., 2024). We converted erosion rates from units of length per time to mass per area per time using bulk density values inferred for the N_{surface} sample (Table S2). We refer to the flux values corrected for inheritance only (F_{met}) as inheritance-corrected fluxes and the inheritance- and erosion-corrected fluxes ($F_{\text{cor}_\varepsilon}$) as erosion-corrected fluxes.

We used gridded datasets and univariate regression to assess the correlation between F_{met} (and $F_{\text{cor}_\varepsilon}$) and environmental variables: elevation, mean annual precipitation (MAP), snow depth, and snow water equivalent (SWE) (Table 2). The influence of elevation was evaluated using 3-meter resolution LiDAR data (Goulden et al., 2020). We determined MAP at each site using downscaled ($400 \times 400 \text{ m}$) Parameter-elevation Regressions on Independent Slopes Model (PRISM; Daly et al., 2008) data from 2008–2019 (Mital et al., 2022), and snow data products derived from the Airborne Snow Observatory (ASO; Painter et al., 2018a, b). ASO combines airborne LiDAR and imaging spectrometry with snow energy balance models to generate spatially continuous snow data. Snow depth is calculated as the difference between a snow-covered LiDAR survey and a snow-free “bare-earth” LiDAR scan. For this study, we used ASO-derived snow depth data (3 m resolution) acquired on March 31, 2018, and April 7, 2019 (Painter et al., 2018a; Deems, 2019). We used SWE data (50 m resolution) from the same dates with snow depth (Painter et al., 2018b). We used the coefficient of determination (R^2) from linear regression analysis to assess the degree variability in F_{met} and $F_{\text{cor}_\varepsilon}$ is explained by elevation and precipitation metrics. The significance of each regression model was determined using the associated p-value. For the precipitation-based metrics, we additionally calculated regression relationships with the y-intercept forced to be zero, because F_{met} and $F_{\text{cor}_\varepsilon}$ should be zero where precipitation is zero, assuming dry fallout of $^{10}\text{Be}_{\text{met}}$ is negligible. We also predicted F_{met} for each of the five calibration sites using the precipitation- and latitude-based empirical model of Graly et al. (2011) using the down-scaled PRISM MAP data (Mital et al., 2022). We re-

scaled the Graly et al. (2011) predictions, which are based on present-day precipitation measurements, to Holocene-averaged values by multiplying them by 1/0.67 to account for differences in paleomagnetic and solar intensity variations (e.g., Clow et al., 2020; Deng et al., 2020) and compared the re-scaled predictions against our F_{met} and F_{cor_e} values.

Using the linear regression parameters, we used the gridded elevation and precipitation datasets to generate spatially continuous predictions of F_{met} and F_{cor_e} across the East River watershed, including the Washington Gulch and Slate River tributaries. We used the regression relationships with a forced y-intercept of zero to generate predictions based on MAP, snow depth, and SWE to ensure model predictions were physically plausible (e.g., did not predict any negative fluxes). We calculated the watershed-averaged flux predicted by elevation, MAP, snow depth, and SWE and compared those predictions against GCM-based predictions, specifically the pre-industrial flux predicted by Heikkilä and von Blanckenburg (2015) and the GEOS-Chem 14.1.1 and ECHAM6.3-HAM2.3 model predictions of Zheng et al. (2024). The Heikkilä and von Blanckenburg (2015) predictions are based on the ECHAM5-HAM model and use a solar modulation factor (ϕ) of 501.76 MeV that averages over three recent decades, but are adjusted to Holocene-average values. The Zheng et al. (2024) predictions use a ϕ value of 488 MeV and average over the period from 2005–2013. We multiplying the Zheng et al. (2024) predicted fluxes by 1.23 (e.g., Heikkilä and von Blanckenburg, 2015), which approximately re-scales them to Holocene-averaged values.

4 Results

4.1 Soil pH

Soil pH for the uppermost sample interval of each profile ranged from 4.3–6.3 and generally increased with depth (Figure 3; Table S3). Mean pH values averaged across all depths for Copper Creek, Washington Gulch North, Gothic, Washington Gulch South, and East River Terminal moraines are 4.7, 5.6, 5.7, 5.9, and 6.3, respectively. The most acidic soils are found at the Copper Creek moraine site, which is the highest elevation and is dominated by spruce-fir forest vegetation (Figure S1), where the lowest pH value of 4.3 was measured in the 0 to 10 cm depth increment.

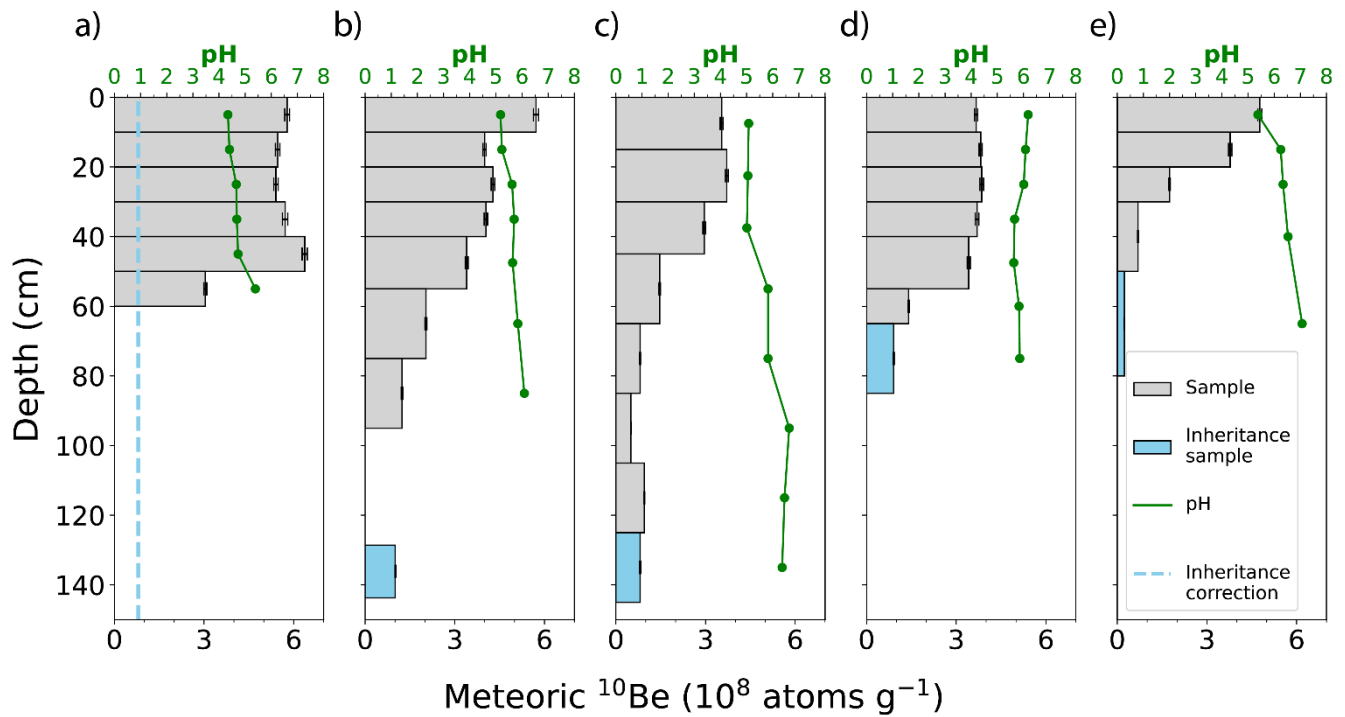


Figure 3. $^{10}\text{Be}_{\text{met}}$ concentration and pH as a function of depth for (a) Copper Creek, (b) Washington Gulch North, (c) Gothic, (d) Washington Gulch South, and (e) East River Terminal moraines. Grey boxes show measured concentrations and uncertainties. Blue boxes indicate samples assumed to represent the inherited $^{10}\text{Be}_{\text{met}}$ concentration. In (a), the blue dashed line shows the mean inheritance value calculated from profiles (b)–(e). The sample used for the inheritance correction in (b) was sampled from a landslide exposure on the Washington Gulch North moraine that was 374 m away from the soil pit where the other samples were collected.

4.2 $^{10}\text{Be}_{\text{met}}$ profiles, inventories, fluxes, and controls on fluxes

$^{10}\text{Be}/^9\text{Be}$ ratios typically were on the order of 1.0×10^{-12} to 10×10^{-12} , which is several orders of magnitude greater than the 2×10^{-15} to 5×10^{-15} ratios measured in blanks (Table 2). Measured $^{10}\text{Be}_{\text{met}}$ concentrations ranged from about 2.4×10^7 to $6.4 \times 10^8 \text{ atom g}^{-1}$ (Table 2). The $^{10}\text{Be}_{\text{met}}$ concentration profiles exhibit a variety of shapes. Profiles at four of the sites have relatively uniform concentrations in the upper $\sim 0.5 \text{ m}$ of the profile, with maximum concentrations at or near the ground surface that decline at greater depths (Table 2; Figure 3a-d). At the lowest elevation site on the terminal moraine, the $^{10}\text{Be}_{\text{met}}$ concentration is greatest at the surface and declines exponentially with depth (Figure 3e). The shallowest profile was at the Copper Creek site, where a coarse, indurated stone layer 60 cm below the ground surface prevented deeper sampling. At Copper Creek the concentration

profile is well mixed in the upper 40 cm, exhibits a subsurface peak concentration 40-50 cm below the ground surface, and then declines sharply in the 10 cm interval directly above the stone line, but not to levels as low as those observed in the deepest samples from other profiles (Figure 3a). Inheritance-corrected $^{10}\text{Be}_{\text{met}}$ inventories range from 7.68×10^9 to 2.35×10^{10} atoms cm^{-2} (Table 3), with the largest and smallest inventories at the Copper Creek and East River terminal moraine sites, respectively. Inheritance-corrected F_{met} values range from 0.43×10^6 to 1.80×10^6 atom $\text{cm}^{-2} \text{yr}^{-1}$ (Table 3). Correcting for inheritance, erosion, and decay using eq. 3 yields F_{cor_e} values of 0.86×10^6 to 1.81×10^6 atom $\text{cm}^{-2} \text{yr}^{-1}$ (Table 3).

The potential impact of erosion on fluxes varies among the sites based on both the inferred erosion rate and the $^{10}\text{Be}_{\text{met}}$ concentration at the soil surface (Figure 4). We inferred erosion rates for the Copper Creek, Washington Gulch North, Washington Gulch South, and East River Terminal sites were 0.0, 0.017, 0.006, and 0.012 mm yr^{-1} respectively. The curvature value of zero at the Copper Creek moraine predicts there is no surface lowering due to erosion, implying the erosion-corrected flux is the same as the measured flux, though the erosion-corrected value is slightly lower because it also accounts for ^{10}Be decay. Accounting for erosion at the other sites leads to erosion-corrected fluxes that are 1–2 times greater than the measured fluxes (Table 3).

F_{met} values are well-correlated with elevation and precipitation metrics in the East River watershed (Figure 5). F_{met} values are linearly correlated with elevation ($R^2=0.91$), mean annual precipitation ($R^2=0.82\text{--}0.85$), mean snow depth in 2018–2019 ($R^2=0.95\text{--}0.97$), and mean SWE in 2018–2019 ($R^2=0.97\text{--}0.98$). Due to the large range of elevation and precipitation within the watershed (Figure 2), the regression models predict substantial within-watershed variability in F_{met} (Figure 6). The erosion-corrected (F_{cor_e}) fluxes exhibit similar trends with the environmental variables, but the correlations are weaker, with R^2 values of 0.43 for elevation, 0.46 to 0.63 for mean annual precipitation, -0.42 to 0.58 for mean snow depth in 2018–2019, and -0.07 to 0.55 for mean SWE in 2018–2019 (Figure 5). Negative R^2 values for the centered regression models forced to have a y-intercept of zero indicate the mean is a better predictor than the regression relationship.

Table 2. $^{10}\text{Be}_{\text{met}}$ sample information and AMS results.

Site	Sample ID ^a (LLNL BE number)	Sample mass (g)	⁹ Be added as carrier (μg)	$^{10}\text{Be}:$ ⁹ Be ratio ($\times 10^{-12}$)	$^{10}\text{Be}:$ ⁹ Be ratio uncertainty ($\times 10^{-14}$)	^{10}Be concentration and uncertainty ^b (10^3 atoms g^{-1})	Inheritance- corrected ^{10}Be concentration and uncertainty (10^3 atoms g^{-1})
Copper Creek	ER_CCT_pit_1_0-10 (BE54737)	0.5227	353.3	12.8	8.50	577993 $\pm$ 7878	502192 $\pm$ 8009
	ER_CCT_pit_1_10-20 (BE54738)	0.5396	352.9	12.5	8.25	546146 $\pm$ 7433	470345 $\pm$ 7572
	R_ER_CCT_Pit_1_20_30 (BE54926)	0.5174	351.7	11.9	8.08	540389 $\pm$ 7405	464588 $\pm$ 7544
	ER_CCT_pit_1_30-40 (BE54740)	0.4994	352.9	12.1	11.5	571221 $\pm$ 8701	495420 $\pm$ 8820

Washington Gulch North	ER_CCT_pit_1_40-50 (BE54741)	0.4998	352.9	13.5	8.95	636818±8676	561017±8795
	ER_CCT_pit_1_50-60 (BE54743)	0.5004	352.2	6.47	7.15	304159±4941	228358±5147
	ER_S2_pit_1_0-10 (BE54729)	0.5456	353.8	13.2	1.16	571850±8461	464670±8682
	ER_S2_pit_1_10-20 (BE54730)	0.5366	353.3	9.09	6.58	399796±5570	292616±5900
	ER_S2_pit_1_20-30 (BE54731)	0.5165	352.2	9.39	6.71	427730±5938	320550±6249
	ER_S2_pit_1_30-40 (BE54732)	0.4948	353.5	8.47	6.13	404217±5631	297037±5958
	ER_S2_pit_1_40-55 (BE54733)	0.4992	353.4	7.19	5.21	339990±4738	232810±5122
	ER_S2_pit_1_55-75 (BE54735)	0.5032	352.4	4.35	4.15	203430±3105	96250±3665
	ER_S2_pit_1_75-95 (BE54736)	0.4959	353.1	2.60	4.82	123570±2725	16390±3349
	ER24_S2_Pit_2_130-145_A (BE55950)	0.5249	351.1	2.40	3.29	107180±1947	0±0
Gothic	ER_GM_pit_1_0-15 (BE54756)	0.4996	352.3	7.52	4.88	354209±4802	273036±5124
	ER_GM_pit_1_15-30 (BE54758)	0.4975	350.6	7.89	4.06	371411±4817	290238±5138
	ER_GM_pit_1_30-45 (BE54759)	0.4983	351.9	6.27	5.5	295743±4374	214570±4725
	ER_GM_pit_1_45-65 (BE54760)	0.5025	352.1	3.14	3.65	146885±2446	65712±3030
	ER_GM_pit_1_65-85 (BE54761)	0.5308	354.6	1.82	2.93	90938±1505	9765±2337
	ER_GM_pit_1_85-105 (BE54762)	0.5149	351.3	1.11	1.99	50473±1089	0±0
	ER_GM_pit_1_105-125 (BE54765)	0.5073	352.1	2.06	1.81	95406±1413	14233±2279
	ER_GM_pit_1_125-145 (BE54766)	0.5174	351.7	1.79	3.31	81173±1788	0±0
Washington Gulch South	ER_AM_Pit_1_0_10 (BE54780)	0.4927	353.0	7.66	6.72	366586±5421	275648±5626
	ER_AM_pit_1_10-20 (BE54750)	0.4863	352.6	7.89	5.23	382133±5207	291195±5420
	ER_AM_pit_1_20-30 (BE54751)	0.5008	352.0	8.21	9.03	385467±6248	294529±6427
	ER_AM_pit_1_30-40 (BE54752)	0.5210	351.8	8.20	8.53	369860±5848	278922±6039
	ER_AM_pit_1_40-55 (BE54753)	0.5648	351.7	8.22	7.34	341913±5088	250975±5306
	ER_AM_pit_1_55-65 (BE54754)	0.5088	352.0	3.05	3.07	140864±2197	49926±2663
	ER_AM_pit_1_65-85 (BE54755)	0.5308	354.6	2.04	2.34	90938±1505	0±0
East River Terminal	ER_BCTM_pit_1_0-10 (BE54744)	0.4968	351.1	10.1	7.34	476832±6650	452918±6671
	ER_BCTM_pit_1_10-20 (BE54745)	0.4980	352.2	8.00	7.02	377931±5589	354017±5614
	ER_BCTM_pit_1_20-30 (BE54746)	0.5889	353.0	4.37	3.72	174923±2561	151009±2615
	ER_BCTM_pit_1_30-50 (BE54747)	0.4999	353.0	1.47	1.65	69226±1135	45312±1253
	ER_BCTM_pit_1_50-80 (BE54748)	0.4902	352.9	0.500	9.27	23914±530	0±0

^aValues at the end of each sample ID refer to the sample depth interval (cm). ^bReported ¹⁰Be concentrations are corrected using ¹⁰Be:⁹Be ratios measured in process blanks. ¹⁰Be:⁹Be ratios measured in process blanks in the same batches as the samples were: 2.44, 1.64, 2.55, 2.93, 2.11, 2.39, 4.71, 2.92, and 2.70×10^{-15} .

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Table 3: Environmental variable values and ¹⁰Be_{met} fluxes at each calibration site.

Site	Elevation ^a (m)	MAP ^b (mm yr ⁻¹)	Mean snow depth ^c 2018 – 2019 (m)	Mean SWE ^d 2018 – 2019 (m)	Inheritance- corrected ¹⁰ Be _{met} inventory ±1σ uncertainty (10 ¹⁰ atoms cm ⁻²)	Inheritance- corrected ¹⁰ Be _{met} flux ±1σ uncertainty (10 ⁶ atoms cm ⁻² yr ⁻¹)	Inheritance- and erosion-corrected ¹⁰ Be _{met} flux ±1σ uncertainty (10 ⁶ atoms cm ⁻² yr ⁻¹)
Copper Creek	3476	1353	1.90	0.64	2.35±0.64	1.80±0.50	1.81±0.50
Washington Gulch North	2944	784	1.01	0.34	1.80±0.57	1.01±0.32	1.57±0.49
Gothic	2885	885	0.90	0.33	1.27±0.47	0.905±0.339	1.79±0.67
Washington Gulch South	2864	655	0.80	0.31	1.73±0.34	0.998±0.197	1.16±0.23
East River Terminal	2764	585	0.42	0.19	0.768±0.131	0.431±0.074	0.861±0.147

^a elevation (m) from Goulden et al. (2020), ^b mean annual precipitation (MAP) from Mital et al. (2022), (c) mean snow depth (m) from 2018-2019 (Painter, 2018a), and (d) Mean snow water equivalent (SWE) (m) from 2018-2019 (Painter et al., 2018b).

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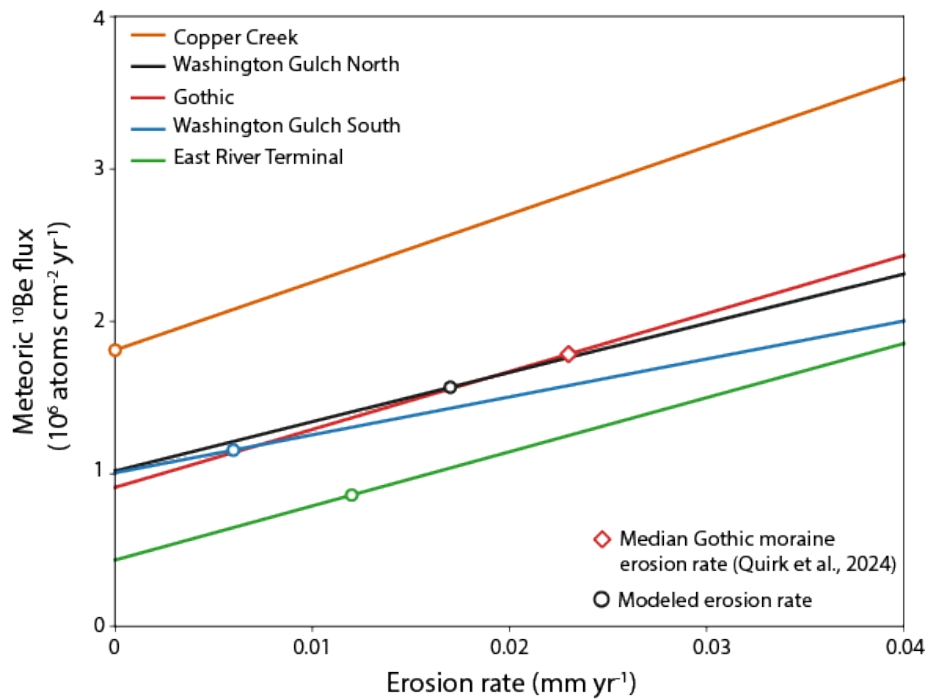


Figure 4. Meteoric ^{10}Be flux as a function of erosion rate. The y-intercept is the measured flux based on soil inventories and moraine ages, which does not account for erosion or decay (decay has a negligible impact on flux). The diamond indicates the measured erosion rate for the Gothic site (Quirk et al., 2024). The circles are erosion rates modelled for each site using equation 4. Predictions for each site have different slopes due to differences in the concentration of $^{10}\text{Be}_{\text{met}}$ at the soil surface.

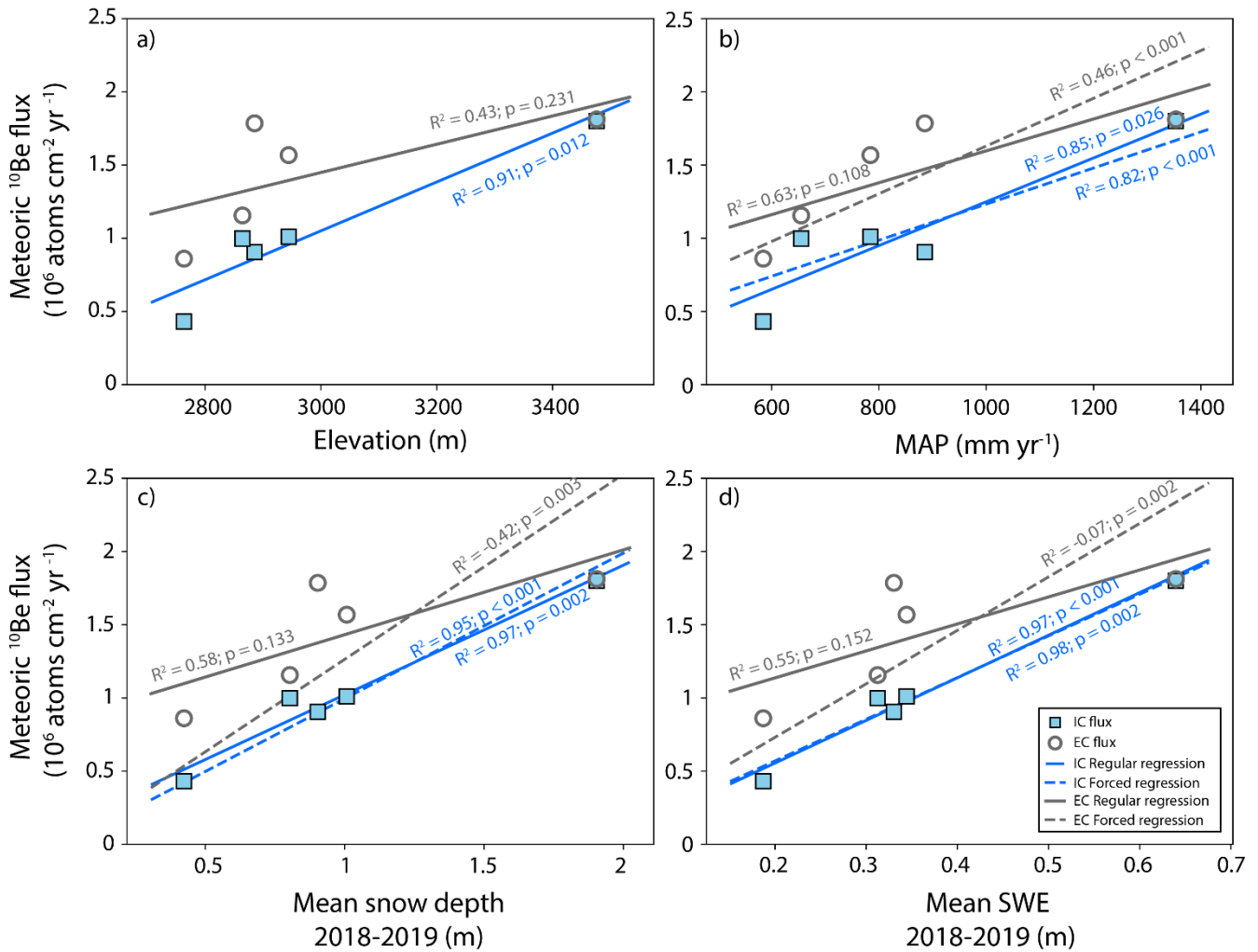


Figure 5. Linear regression models of $^{10}\text{Be}_{\text{met}}$ flux versus (a) elevation, (b) mean annual precipitation (MAP), (c) mean snow depth (2018-2019), and (d) mean snow water equivalent (SWE) (2018-2019). Blue squares show inheritance-corrected fluxes and open gray circles show erosion-corrected fluxes. For the inheritance-corrected (IC) data, solid blue lines show results from standard regression models and the dashed blue lines show regression models where the y-intercept is forced to zero. For the erosion-corrected (EC) data, solid gray lines show standard regression model results and gray dashed lines show regression models where the y-intercept is forced to zero. Forced-zero regressions are shown only for MAP, mean snow depth, and mean SWE. Regression coefficients, R^2 values, and p values are reported in Table S4.

The different regression models predicted watershed-averaged F_{met} values that vary by more than a factor of three (Table 4). For elevation, the regression model yields average F_{met} and F_{cor_e} of 1.31×10^6 and $1.60 \times 10^6 \text{ atoms cm}^{-2} \text{ yr}^{-1}$, respectively. The

MAP-based regression resulted in average F_{met} and F_{cor_e} values of 0.66×10^6 and 1.17×10^6 atoms $\text{cm}^{-2} \text{yr}^{-1}$ for the standard and 0.75×10^6 and 0.99×10^6 atoms $\text{cm}^{-2} \text{yr}^{-1}$ for the forced regression models, respectively. Snow depth models produced average F_{met} and F_{cor_e} values of 1.76×10^6 and 1.91×10^6 and 1.82×10^6 and 2.31×10^6 atoms $\text{cm}^{-2} \text{yr}^{-1}$ for the standard and forced regression models, respectively. Similarly, SWE-based regressions yielded watershed-averaged F_{met} and F_{cor_e} values of 1.42×10^6 and 1.68×10^6 for the standard regression and 1.41×10^6 and 1.81×10^6 atoms $\text{cm}^{-2} \text{yr}^{-1}$ for the forced regressions, respectively. We assessed the influence of extreme values and potential outliers by calculating watershed-averaged F_{met} values using only pixels with the lowest 99th, 95th, and 90th percentile values, which resulted in lower mean F_{met} values, with the greatest reductions for the precipitation- and snow-based estimates (Table 4).

Table 4. Watershed-averaged $^{10}\text{Be}_{met}$ flux predictions based on all pixels (100%) and when excluding pixels with high values by using only the lowest 99, 95, and 90th percentile values when calculating the mean.

Variable	Regression type	Meteoric ^{10}Be flux (10^6 atoms $\text{cm}^{-2} \text{yr}^{-1}$)							
		100%		99%		95%		90%	
		Inheritance-corrected	Erosion-corrected	Inheritance-corrected	Erosion-corrected	Inheritance-corrected	Erosion-corrected	Inheritance-corrected	Erosion-corrected
Elevation (m)	Regular	1.31	1.60	1.30	1.59	1.25	1.56	1.20	1.53
MAP (mm yr^{-1})	Regular	0.66	1.17	0.64	1.16	0.60	1.12	0.55	1.09
	Forced	0.75	0.99	0.73	0.97	0.70	0.92	0.65	0.86
Mean snow depth (m) 2018 - 2019	Regular	1.76	1.91	1.70	1.88	1.58	1.80	1.48	1.73
	Forced	1.82	2.31	1.76	2.24	1.62	2.06	1.51	1.91
Mean SWE (m) 2018- 2019	Regular	1.42	1.68	1.37	1.65	1.25	1.58	1.15	1.51
	Forced	1.41	1.81	1.37	1.75	1.25	1.60	1.15	1.47

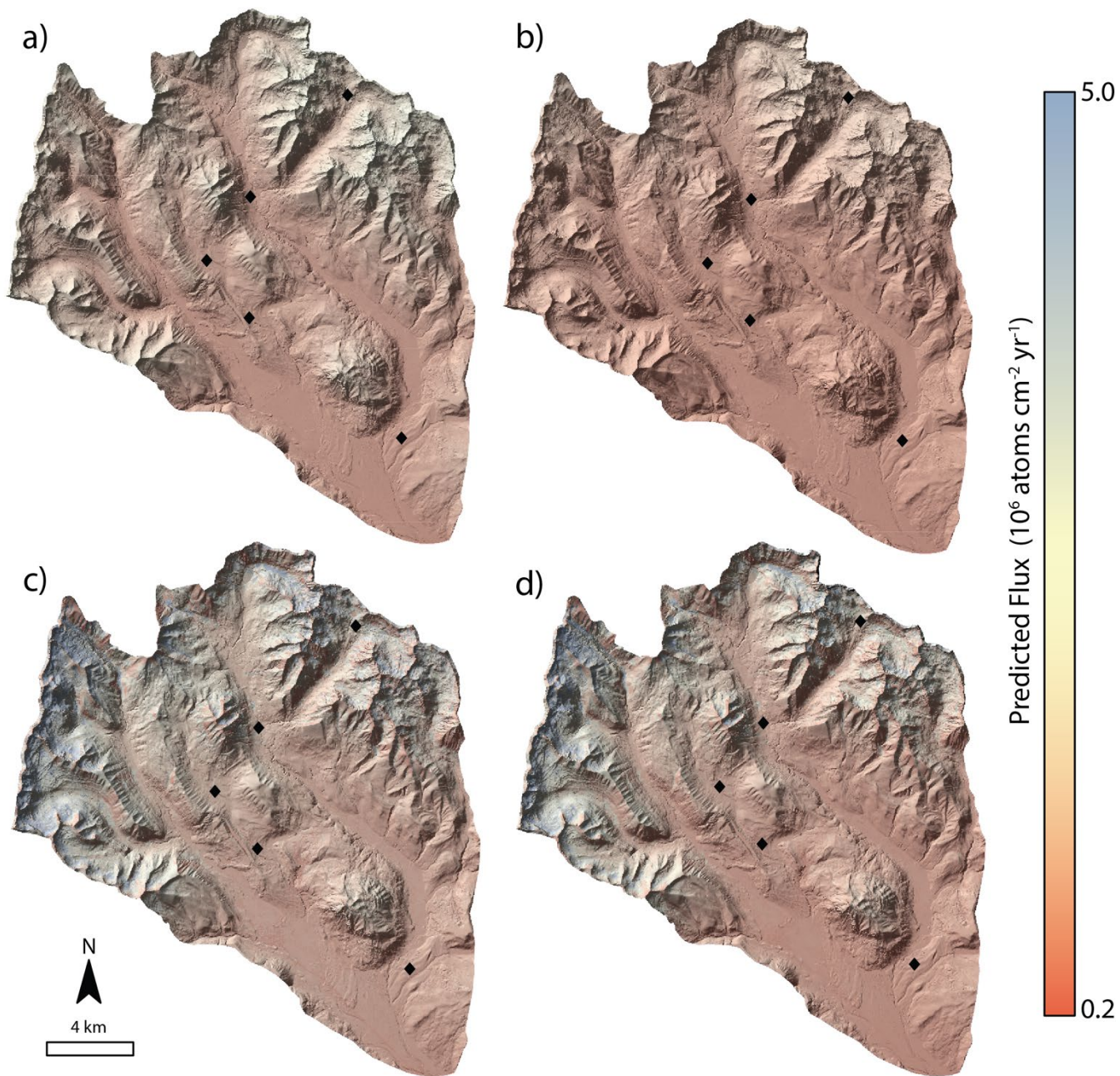


Figure 6. Spatial predictions of $^{10}\text{Be}_{\text{met}}$ flux based on regression models using: (a) elevation; (b) mean annual precipitation; (c) mean snow depth from 2018–2019; and (d) mean snow water equivalent from 2018–2019 based on regression models for inheritance-corrected fluxes. The equivalent results for inheritance- and erosion-corrected fluxes are shown in Figure S2.

5 Discussion

¹⁰Be_{met} profiles tend to have characteristic shapes, and the shapes are sometimes used to infer processes of ¹⁰Be_{met} mobility (e.g., Graly, 2010). The two most common profile shapes are characterized by either a subsurface peak in concentration or concentrations that decline with depth from a maximum value at the soil surface (Graly, 2010). The profile on the East River Terminal moraine exhibits an exponential decline in concentration with depth, however the concentrations at the other four sites are fairly uniform with depth. The uniform profiles likely arise from bioturbation, via vegetation or burrowing mammals. Pocket gophers (*Thomomys talpoides*) are the main soil disturbing mammal within the East River watershed, and soil disturbance can be intense, with surveys indicating soil disturbance by burrowing can exceed one disturbance per m² within the elevation range of our sampling sites (Lynn et al., 2018). The distinct stone line at the Copper Creek site may have been generated by bioturbation (e.g., Johnson, 1989), though we cannot rule out the possibility of a periglacial origin (Ball, 1967). The stone line prevented deeper sampling at the Copper Creek site, which makes it difficult to determine whether the full ¹⁰Be_{met} inventory was sampled, or whether ¹⁰Be_{met} mobilized and is below the stone line. In either case, the high-elevation Copper Creek site has the highest inventory, which is consistent with the higher precipitation rates at the site.

Pedogenic loss of ¹⁰Be_{met} has been documented or suspected in low pH soils with substantial chemical alteration (e.g., Pavich et al., 1984; Bacon et al., 2012; Dixon et al., 2018). The partition coefficient for ¹⁰Be_{met} is high when pH is greater than ~4.5 (You et al., 1989) but declines at lower pH values. Analysis of a compilation of soil ¹⁰Be_{met} and pH data suggest loss is most likely to occur below a pH of 3.9 (Graly et al., 2010). The 0-10 cm interval sample at Copper Creek site had the lowest measured pH value of 4.3, but pH increased with depth to a value of 5.4 at the base of the soil profile, and pH values for the other sites are higher. Hence we infer ¹⁰Be_{met} loss via chemical leaching is minimal and that mobilized ¹⁰Be_{met} is likely confined within the sampled horizons. Minimal chemical loss of ¹⁰Be_{met} was also inferred for moraines near Pinedale, Wyoming, USA, which has a climate and ecosystem broadly similar to the East River (Clow et al., 2020).

Glacial moraines erode via diffusive soil transport processes (e.g., Putkonen and Swanson, 2003). The F_{met} values we measure are hence minimum values because erosional loss of ¹⁰Be_{met} reduces inventories. Incorporating the median erosion rate for the Gothic Moraine inferred from the ¹⁰Be_{in situ} profile (Quirk et al., 2024) causes the flux inferred for that site to double. However, the other sites have lower convexity than the Gothic site (Table 1), which reduces the influence of erosion. The Copper Creek site has a planar topography (curvature = 0 m⁻¹), and hence the modelled erosion rate is zero. Accounting for topographic influences on erosion rates at the other three sites predicts that fluxes are 1.2 to 2.0 times greater than the values that are not corrected for erosion. Although the diffusion coefficient we calculate is comparable to values measured in other landscapes

with similar climates (Richardson et al., 2019), there is still inherent uncertainty in using modelled, rather than measured erosion rates, to correct the measured fluxes.

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F_{met} is highly correlated with elevation, MAP, snow depth, and SWE, which is unsurprising because these variables are related to one another via orographic controls on precipitation rates. However, different slope and intercept values for regression relationships result in different F_{met} predictions for the same location and watershed-averaged fluxes that differ by a factor of about 2 to 3. The precipitation data average over only a few decades and snow depth and SWE datasets cover only two winters, which are only small fractions of the post-glacial period over which $^{10}\text{Be}_{met}$ has accumulated in soils. The recent snowfall magnitude and spatial patterns may also differ from Holocene averages. However, unlike elevation and MAP datasets, the snow-based inventories yield insight on how local-scale topography may influence spatial patterns of F_{met} . For example, we predict that lower snow accumulation on sharp ridges relative to adjacent slopes leads to lower $^{10}\text{Be}_{met}$ accumulation on ridges (Figure 6). Watershed-averaged fluxes predicted using snow depth and SWE are more sensitive to extreme values than averages based on elevation (Table 4). The mean snow depth of 1.90 m at the highest elevation Copper Creek site is a fraction of the highest snow depth values, which exceed 20 m (Figure S3) and are potentially artefacts. The lack of F_{met} calibration sites in areas with higher snow accumulation may result in inaccurate predictions for such areas, however the lack of moraines at higher elevations makes it difficult to measure $^{10}\text{Be}_{met}$ fluxes in such areas. The MAP data are based on decadal-scale precipitation measurements but are interpolated based on elevation. However, the watershed-averaged MAP-based predictions are notably lower than values based on elevation and snow data, suggesting precipitation may be under-estimated. Elevation is the only parameter not subject to temporal bias. Hence elevation should be a robust predictor of $^{10}\text{Be}_{met}$ fluxes, though elevation-based predictions may be biased where local topography leads to persistent snow redistribution via wind and avalanches. Given the different $^{10}\text{Be}_{met}$ flux predictions from the elevation, MAP, snow depth, and SWE datasets, an ensemble approach may be appropriate for using $^{10}\text{Be}_{met}$ values to quantify surface process rates, in a manner similar to assessing results from multiple production rate scaling models when determining $^{10}\text{Be}_{in situ}$ exposure ages (e.g., Balco et al., 2008).

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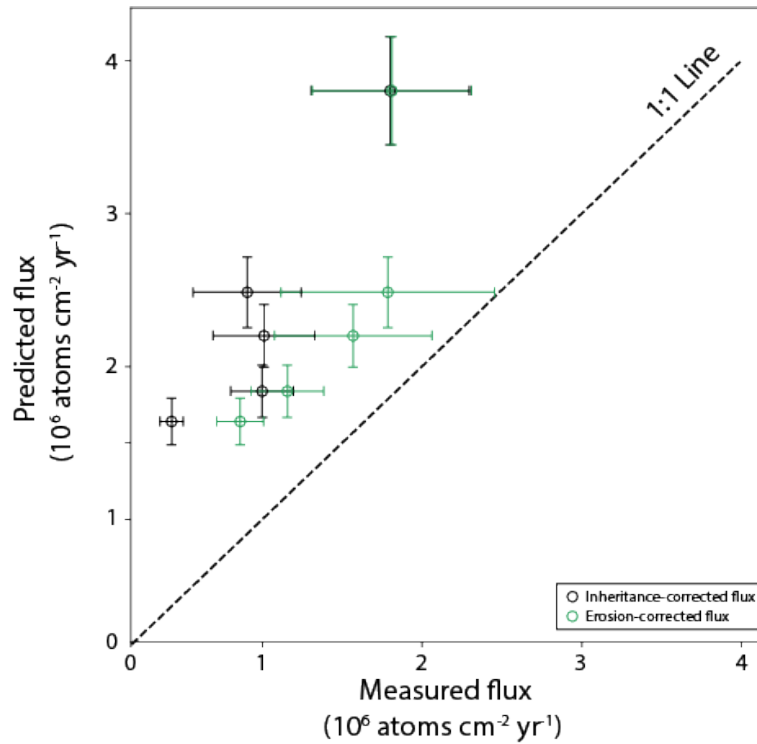


Figure 7. Predicted $^{10}\text{Be}_{\text{met}}$ flux from the empirical precipitation-based model of Graly et al. (2011), scaled to reflect Holocene-average values (e.g., Clow et al., 2020) versus measured flux values for the five study sites.

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Fluxes predicted by an empirical precipitation- and latitude-based model (Graly et al., 2011) are 1.8–3.8 times higher than measured inheritance-corrected fluxes (Figure 7). The erosion-corrected fluxes are closer to the empirical predictions, but are still 1.4–2.1 times higher than observed values. The differences between the predicted and measured point-scale values are generally similar to watershed-average predictions, although GCM-based models both over- and under-predict fluxes. GCM-

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based models predict $^{10}\text{Be}_{\text{met}}$ fluxes of $1.46 \times 10^6 \text{ atoms cm}^{-2} \text{ yr}^{-1}$ (Heikillä and von Blanckenberg, 2015, pre-industrial/Holocene-average production), $1.63 \times 10^6 \text{ atoms cm}^{-2} \text{ yr}^{-1}$ (Zheng et al., 2024, ECHAM6.3-HAM2.3), and $2.12 \times 10^6 \text{ atoms cm}^{-2} \text{ yr}^{-1}$ (Zheng et al., 2024, GEOS-Chem 14.1.1) for the East River watershed and its tributaries (Figure 8). A single grid-cell from the GCM-models covers an area larger than the watershed, hence the models simulate the influence of meso-scale atmospheric dynamics on F_{met} , but do not capture within-watershed F_{met} variability caused by orographic precipitation. The measured F_{met} and F_{cor_e}

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values and GCM-based predictions vary by nearly a factor of four for the most extreme case. However, because they cover a larger spatial scale, our empirically-predicted watershed-averaged F_{met} values are more directly comparable to the GCM-based predictions than the point-based measurements on each moraine. The watershed-averaged F_{met} values predicted by snow depth ($1.76\text{--}1.82 \times 10^6 \text{ atoms cm}^{-2} \text{ yr}^{-1}$) are greater than the Heikkilä and von Blanckenburg (2015) and Zheng et al. (2024) ECHAM6.3-HAM2.3 predictions, but lower than the Zheng et al. (2024) GEOS-Chem 14.1.1 prediction (Figure 8). However,

the watershed-averaged F_{met} values predicted by elevation (1.31×10^6 atoms $\text{cm}^{-2} \text{yr}^{-1}$) and SWE ($1.41\text{--}1.42 \times 10^6$ atoms $\text{cm}^{-2} \text{yr}^{-1}$) are similar to the 1.46×10^6 atoms $\text{cm}^{-2} \text{yr}^{-1}$ pre-industrial prediction of Heikkilä and von Blanckenburg (2015) and the 1.63×10^6 atoms $\text{cm}^{-2} \text{yr}^{-1}$ ECHAM6.3-HAM2.3 prediction of Zheng et al. (2024) (Figure 8). The MAP-based predictions of $0.66\text{--}0.75 \times 10^6$ atoms $\text{cm}^{-2} \text{yr}^{-1}$ are lower than all of the GCM-based predictions (Figure 8). Hence the best agreement between GCM-based predictions our empirically-estimated watershed-averaged fluxes are from our elevation- and SWE-based models.

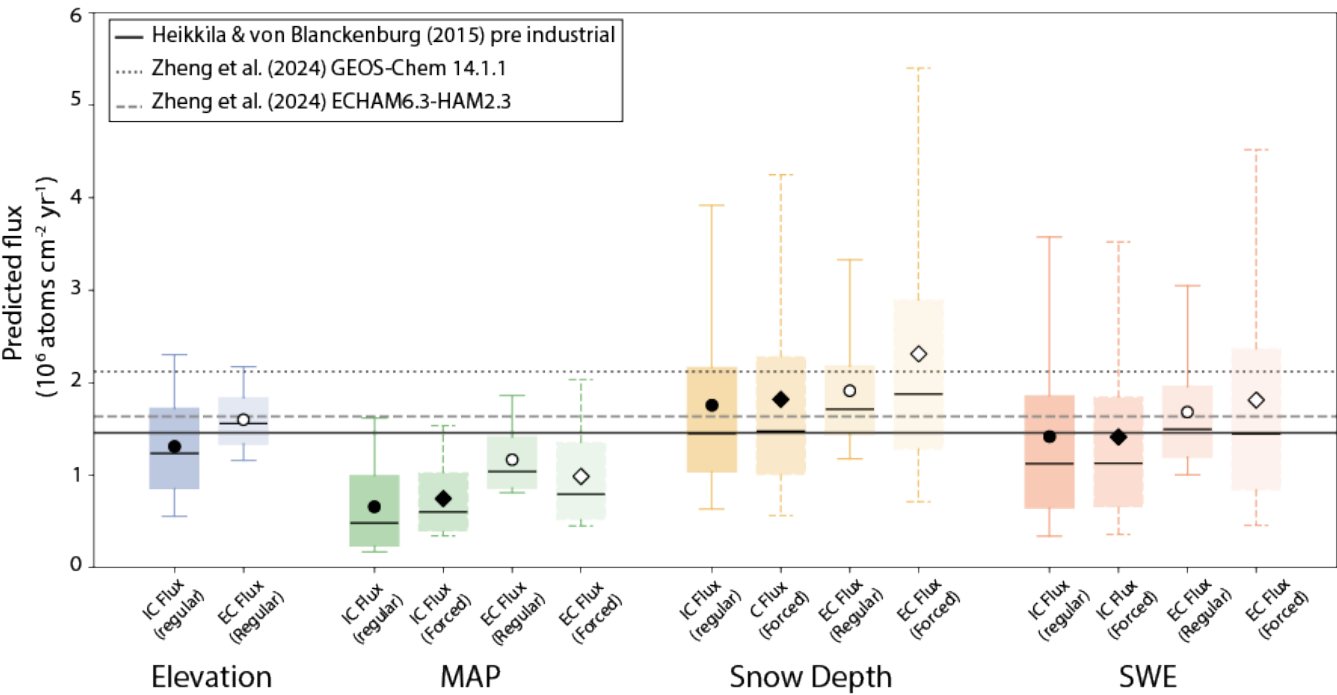


Figure 8: Watershed-scale $^{10}\text{Be}_{met}$ fluxes predicted by empirical models based on elevation, mean annual precipitation (MAP), mean snow depth, and mean snow water equivalent (SWE) and flux predictions from global circulation-based (GCM) models. Boxplots show the median (horizontal line), interquartile range, and 5th and 95th percentiles of flux values. Mean fluxes are show with circles (standard regression models, left box) and diamonds (forced regression models, right box). GCM-based flux predictions from Heikkilä and von Blanckenburg (2015) and Zheng et al. (2024) are shown as horizontal lines. IC indicates inheritance-corrected fluxes and EC indicates erosion-corrected fluxes.

Comparing measured $^{10}\text{Be}_{met}$ fluxes against empirical models generated with data from other regions and GCM-based predictions is somewhat complicated by the likely contribution of dust to the measured values. The contribution of dust to

¹⁰Be_{met} inventories in the East River is unconstrained, but dust fluxes inferred from lacustrine sediment in the nearby San Juan
455 Mountains increased from ~0.01 g cm⁻² yr⁻¹ to 0.4 g cm⁻² yr⁻¹ over the last several centuries (Ballantyne et al., 2011) and recent
snowpack measurements indicate seasonal dust fluxes are on the order of 5-10 g cm⁻² yr⁻¹ (Lawrence et al., 2010). Dust fluxes
in the Colorado Front Range are 0.1–0.2 g cm⁻² kyr⁻¹ (Dethier et al., 2012). Dust fluxes in the Front Range contribute <10% to
the total flux of ¹⁰Be_{met} (Ouimet et al., 2015). If the contribution of dust in the East River is comparable to that in the Front
Range, accounting for dust would not substantially influence the magnitude of discrepancies between GCM-based predictions
460 and our empirically derived F_{met} estimates.

At a regional level, ¹⁰Be_{met} fluxes for the Colorado Front Range based on soil inventories have been estimated to be $1.5 \pm 0.2 \times 10^6$
atoms cm⁻² yr⁻¹ (Ouimet et al., 2015). Soil inventory-based F_{met} values in the Wind River Mountains, Wyoming are
 $1.46 \pm 0.20 \times 10^6$ atoms cm⁻² yr⁻¹ and $1.30 \pm 0.48 \times 10^6$ atoms cm⁻² yr⁻¹ for Pinedale- and Bull Lake-aged moraines, respectively
465 (Clow et al., 2020). The $0.43\text{--}1.8 \times 10^6$ atoms cm⁻² yr⁻¹ F_{met} values and $0.86\text{--}1.81 \times 10^6$ atoms cm⁻² yr⁻¹ F_{cor_e} values we measure
are consistent with those from the Front Range and Wind River Mountains, though those sites are drier, with ~500 mm yr⁻¹ of
mean annual precipitation in the Front Range (Ouimet et al., 2015) and 276 mm yr⁻¹ at the Wind River Mountains sites (Clow
et al., 2020), relative to the 585 to 1353 mm yr⁻¹ precipitation at the five East River study sites. The driest site in the East River
(terminal moraine, 585 mm yr⁻¹ MAP) has inheritance- and erosion-corrected ¹⁰Be_{met} fluxes of 0.43×10^6 and 0.86×10^6 atoms
470 cm⁻² yr⁻¹, which are lower than fluxes inferred in the Front Range and Wind River Mountains. Hence although the fluxes are
similar, precipitation alone does not explain all of the regional variability.

Local calibration of ¹⁰Be_{met} flux greatly increases the utility of ¹⁰Be_{met} as a tracer of surface processes in the East River
watershed, where there is prevalent shale bedrock that is not amenable to the more common ¹⁰Be_{in situ} method. Our findings
475 show that, whereas ¹⁰Be_{met} flux predictions from GCM-based models may be broadly applicable for large-scale watershed
applications, such models do not capture the high spatial variability in ¹⁰Be_{met} fluxes that is driven by orographic influences
on precipitation. Hence using ¹⁰Be_{met} to quantify physical erosion and weathering rates at the small watershed or soil profile-
scale requires not only local-scale calibration of ¹⁰Be_{met} fluxes, but also a means for predicting ¹⁰Be_{met} fluxes for sites where
elevation and precipitation varies from that of the sites used to for calibration. Our study demonstrates how high-resolution
480 LiDAR elevation and snowpack datasets can be combined with locally calibrated ¹⁰Be_{met} flux measurements to generate
spatially resolved flux predictions, which furthers our understanding of the processes that control the delivery of ¹⁰Be_{met} to
soils.

485 6 Conclusions

Meteoric ^{10}Be fluxes calculated from soil inventories on five glacial moraines dated to 13.0–17.9 ka in the East River watershed, Colorado, USA range from $0.43\text{--}1.8\times 10^6$ atoms $\text{cm}^{-2}\text{yr}^{-1}$ when corrected for inheritance and $0.86\text{--}1.81\times 10^6$ atoms $\text{cm}^{-2}\text{yr}^{-1}$ after also taking erosional loss of ^{10}Be into account. Soil pH values are above the threshold where loss of meteoric ^{10}Be via leaching is likely to occur. The inheritance-corrected and erosion-corrected fluxes vary systematically with
490 environmental gradients across the watershed and are generally correlated with elevation, mean annual precipitation, mean snow depth, and mean snow water equivalent, reflecting the important role orographic precipitation plays in influencing spatial patterns of meteoric ^{10}Be flux within the watershed. Elevation is the only predictor of flux that is not subject to temporal bias, whereas the snow-based predictions are based on only two years of measurements, and the magnitude and pattern of snow accumulation during those years may vary substantially from Holocene averages. However, the snow-based predictions yield
495 insight into smaller-scale topographic controls on meteoric ^{10}Be fluxes, such as predictions of lower fluxes on sharp ridge crests relative to adjacent hillslopes due to snow redistribution. The fluxes measured at the five moraines are higher than predictions from an empirical model relating flux and mean annual precipitation and latitude (Graly et al., 2011). Watershed-averaged fluxes predicted from empirical correlations between measured flux and elevation, mean annual precipitation, mean snow depth, and mean snow water equivalent are comparable to predictions from global-scale general circulation models.
500 Although the GCM models do not capture the spatial variation in flux within the watershed and cannot be used for soil profile-based applications, the GCM results provide useful estimates of watershed-scale flux. The findings highlight the large degree which meteoric ^{10}Be fluxes vary within alpine landscapes and demonstrate how combining soil meteoric ^{10}Be inventories and high-resolution gridded datasets can generate spatially-resolved meteoric ^{10}Be flux predictions. Our results enhance the utility of meteoric ^{10}Be as a tracer of surface processes in the East River watershed, and other alpine landscapes with rocks that are
505 not conducive to using *in situ*-produced ^{10}Be and add to our overall understanding of the factors that influence the delivery and retention of meteoric ^{10}Be in soils.

7 Data availability

Data are available at the U.S. Department of Energy ESS-DIVE archive: Marmolejo-Cossío J; Larsen I; Marder, E; Hidy A
510 (2025): Meteoric ^{10}Be Flux Calibration Data for the East River Watershed, Colorado, USA; doi:10.15485/3025240

8 Author contribution

JM: data curation, formal analysis, investigation, writing (original draft preparation)
IL: conceptualization, funding acquisition, investigation, methodology, project administration, resources, supervision, writing (original draft preparation, reviewing and editing)
515 EM: formal analysis, data curation, writing (reviewing and editing)
AH: formal analysis, writing (reviewing and editing)

9 Competing interests

The authors declare that they have no conflict of interest.

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