

Geochemistry, Geophysics, Geosystems®



RESEARCH ARTICLE

10.1029/2024GC012020

Key Points:

- Paleointensity, paleolatitude, and atmospheric changes are included in cosmogenic nuclide production rate calculations extending to 70 Ma
- Paleointensity variations from 0 to 70 Ma affect production rates more than paleolatitude or climate-driven changes in atmospheric depth
- Changes in production rates using this model framework will be most significant for pre-Quaternary paleoexposures

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Citation:

Mijjum, M., Bristol, K. E., Bono, R. K., Sprain, C. J., Lifton, N., & Tremblay, M. M. (2025). A model framework for scaling pre-quaternary cosmogenic nuclide production rates. *Geochemistry, Geophysics, Geosystems*, 26, e2024GC012020. <https://doi.org/10.1029/2024GC012020>

Received 7 NOV 2024

Accepted 20 DEC 2024

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A Model Framework for Scaling Pre-Quaternary Cosmogenic Nuclide Production Rates

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Abstract Cosmogenic nuclide dating is an essential component of studying Earth surface processes, but it requires knowledge of how nuclide production rates vary in time and space. Typically, production rates are calibrated at sites with independently well-constrained exposure histories and then scaled to other sites of interest using scaling frameworks that account for spatial and temporal variations in the secondary cosmic-ray flux at Earth's surface. To date, scaling schemes for terrestrial cosmogenic nuclide production rates have been developed for the Quaternary, yet cosmogenic nuclide applications that extend beyond the Quaternary are becoming more prevalent. For these deeper time applications, production rate calculations using scaling models optimized for the latest Quaternary neglect longer term spatiotemporal variations in geomagnetic field intensity, paleogeography, and paleoatmospheric depth. We present a production rate scaling scheme for the past 70 million years, SPRITE (Scaling Production Rates In deep TimE). This framework extends existing scaling schemes into deeper time by (a) accounting for site-specific changes in paleolatitude, (b) integrating a geomagnetic field intensity model rooted in data from a global paleomagnetic database, and (c) incorporating climate-driven, time-varying atmospheric depths. We evaluate the efficacy of our model by applying it to existing data sets from paleoexposure sites, and from sites with apparent continuous million-year exposure histories. This scaling model can be applied with measurements of stable cosmogenic nuclides to research questions such as constraining hiatus durations between ancient lava flows and calculating the formation timescales of stable landforms in arid environments over millions of years.

Plain Language Summary Dating geologically old surfaces (older than a few million years) is useful for understanding climatic, tectonic, and volcanic processes. To quantify these timescales, we can use cosmogenic nuclides, which are produced in rocks at the Earth's surface via interactions with high energy particles ultimately originating from the galaxy called cosmic rays. Earth's magnetic field deflects these charged particles, so as the magnetic field intensity changes over time and space, the production rate of cosmogenic nuclides changes as well. Over millions of years, the positions of geologic surfaces relative to the Earth's magnetic field also change due to the motion of tectonic plates. Changes in climate will also affect how thick the atmosphere is locally and therefore the interactions of cosmic rays with the atmosphere before they reach Earth's surface. In this work, we create a new model framework to calculate cosmogenic nuclide production rates over the last 70 million years that accounts for the effects of changes in the magnetic field strength, the location of tectonic plates, and atmospheric thickness over this timespan. Incorporating these effects impacts calculated exposure ages of ancient geologic surfaces and therefore our interpretations of the geologic history of these surfaces from cosmogenic nuclide measurements.

1. Introduction

Measurements of in situ cosmogenic nuclides in minerals have become an essential component of quantitatively studying Earth surface processes (e.g., reviews by Granger et al., 2013; Schaefer et al., 2022). High-energy primary galactic cosmic rays (GCRs, dominantly protons) incident on the Earth's atmosphere induce a cascade of secondary cosmic-ray particles (such as neutrons, protons, and muons) via interactions with atmospheric nuclei. These secondary particles can yield nuclear reactions with target nuclei in the atmosphere to generate “meteoric” cosmogenic nuclides, or if they have high enough energies, they can reach the surface and produce in situ cosmogenic nuclides within near-surface minerals. In situ cosmogenic nuclides are primarily produced via spallation reactions between high-energy neutrons or protons and target elements present in minerals (e.g., Si, O,

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Mg, Al), and are continuously produced while a rock or mineral is exposed within the uppermost few meters of the Earth's surface (e.g., Granger et al., 2013). Production via spallation decreases approximately exponentially with depth, with typical attenuation lengths of 160 g/cm² (Gosse & Phillips, 2001), such that production via spallation becomes negligible at depths greater than ~3 m (e.g., Bierman et al., 2002; Gosse & Phillips, 2001). Cosmogenic nuclides are also produced via interactions with muons, which have much longer attenuation lengths of several thousand g/cm² (e.g., Heisinger et al., 2002a, 2002b), but also have a much lower likelihood of producing cosmogenic nuclides than high-energy neutrons or protons. Muogenic production accounts for <~2% of the total surface cosmogenic nuclide production rate for ²¹Ne (Balco et al., 2019) and ~0.1% for ³He (Larsen et al., 2021).

Although many nuclides can be produced by cosmic-ray interactions, there are only a handful of cosmogenic nuclides that are widely utilized for studying the surface exposure histories of rocks. These commonly used cosmogenic nuclides share the characteristic that they are absent or extremely rare in rocks and minerals that have not been exposed to cosmic ray-induced production. Cosmogenic ¹⁰Be and ²⁶Al are the two most commonly utilized cosmogenic nuclides for studying surface exposure histories and surface process rates (e.g., Alvarez-Marrón et al., 2008; Balco et al., 2005; Granger et al., 1997; Hofmann et al., 2022; Nishiizumi et al., 2005; Peri et al., 2022; Vandermaelen et al., 2022; Zhou et al., 2020). As ¹⁰Be and ²⁶Al are radionuclides with relatively short half-lives ($t_{1/2} = 1.39 \times 10^6$ yr and 7.05×10^5 yr, respectively, Chmeleff et al., 2010; Korschinek et al., 2010; Norris et al., 1983), we can be confident that ¹⁰Be and ²⁶Al concentrations measured in rock samples reflect the geologically recent exposure of those rocks to cosmic rays. However, the short half-lives of ¹⁰Be and ²⁶Al also mean that they can only be used to study exposure histories within the last several million years before their concentrations reach secular equilibrium between production and decay. The majority of applications of cosmogenic ¹⁰Be and ²⁶Al are therefore restricted to the Quaternary (e.g., Ancrenaz et al., 2022; Clow et al., 2023; Corbett et al., 2016; Ivy-Ochs et al., 2006; Kim & Sutherland, 2004; Matmon et al., 2005; Yang et al., 2019).

To study longer duration exposure histories or to look at paleoexposure histories that significantly predate the Quaternary, the stable cosmogenic noble gases are useful. The most commonly used cosmogenic noble gases are ³He and ²¹Ne, and we focus on these isotopes in our model framework, recognizing that cosmogenic stable isotopes of krypton are also currently being developed for terrestrial applications (Dunai et al., 2022). Noble gas atoms are rare in rocks and minerals, and although ³He and ²¹Ne can originate from sources other than cosmogenic production (e.g., trapped magmatic or atmospheric sources; Niedermann, 2002), these non-cosmogenic components can often be corrected based on the noble gas isotopic composition and/or by the analysis of samples shielded from cosmic rays (e.g., Kurz, 1986). There are a growing number of studies that utilize cosmogenic noble gases to study the exposure histories of surfaces on timescales longer than the Quaternary (e.g., Bierman & Caffee, 2001; Carrizo et al., 2008; Dunai et al., 2005; Evenstar et al., 2017; Nishiizumi et al., 2007; Ritter, Stuart, et al., 2018; Ritter et al., 2022; Schäfer et al., 1999), or to quantify the paleoexposure histories of surfaces or sediments that were exposed to cosmic rays in the geologic past (e.g., Ben-Israel et al., 2018; Blard et al., 2005; Hofmann et al., 2017; Libarkin et al., 2002; Sinclair et al., 2019).

Motivated by the growing number of cosmogenic noble gas applications that extend beyond the Quaternary, in this paper we introduce a model framework for scaling terrestrial cosmogenic noble gas production rates and calculating exposure ages over the last 70 million years which we call SPRITE, which stands for Scaling Production Rates In deep Time. Production rate scaling frameworks developed to date are optimized for the late Quaternary, both because this is the time period over which most cosmogenic nuclide observations have been applied, and because current full-vector (direction and intensity), globally continuous, and time-varying geomagnetic field models do not extend beyond 100 ka (Panovska et al., 2018). As we discuss below, temporal changes in the Earth's magnetic field affect cosmogenic nuclide production rates, and therefore need to be accounted for in production rate scaling frameworks. The most-recently developed, least biased, and most flexible scaling model in use today is the LSDn model (Lifton-Sato-Dunai nuclide-specific; Lifton et al., 2014) which accounts for differing energy-dependencies of production for different nuclides. For periods beyond the late Quaternary (>2 Ma), current implementations of the LSDn model framework (Lifton, 2016; Lifton et al., 2014) utilize the mean field strength of the GLOPIS-75/PADM2M record for the last 2 Ma (Laj et al., 2004; Ziegler et al., 2011). However, the paleomagnetic record shows substantial variation in the strength of the Earth's magnetic field before 2 Ma (e.g., Biggin et al., 2009, 2015; Bono et al., 2021, 2022; Kulakov et al., 2019; Selkin & Tauxe, 2000), which is unaccounted for at present in LSDn.

In order to improve applications of cosmogenic noble gases beyond the Quaternary, SPRITE extends the current LSDn framework by implementing the Monte Carlo Axial Dipole Average Model (MCADAM), a continuous paleointensity model spanning from 50 ka to 4.2 Ga (Bono et al., 2022), to account for changes in paleomagnetic field strength. SPRITE also incorporates the effects of changes in paleolatitude due to tectonic plate motions and climate-driven changes in atmospheric depth on cosmogenic noble gas production rates over the last 70 Ma. We demonstrate the effects of incorporating changes in the geomagnetic field, paleolatitude, and climate-driven changes in paleoatmospheric depth on pre-Quaternary timescales using hypothetical examples as well as two examples from the cosmogenic nuclide literature. First, we use observations of cosmogenic ^3He in pyroxene and ^{21}Ne in quartz from surfaces in the Atacama Desert (Dunai et al., 2005; Evenstar et al., 2017) to quantify their apparent surface exposure durations extending from the present day to the Oligocene. Second, we use observations of cosmogenic ^{21}Ne in quartz from a depth profile of the Fish Canyon Tuff (Libarkin et al., 2002) to quantify the erosional history of this volcanic deposit before it was buried by the Carpenter Ridge Tuff in the late Oligocene. We use these examples to evaluate the advantages and limitations of SPRITE for studying pre-Quaternary surface exposure histories using cosmogenic noble gases, and describe ways in which this model could be improved in the future.

1.1. Cosmogenic Noble Gas Production Rates

The concentration of a cosmogenic noble gas $C(z, t)$ after some exposure duration t and at a depth below the surface z can be calculated according to Equation 1 after Niedermann (2002):

$$C(z, t) = C(z, 0) + \frac{P_{0,sp}(t)}{\rho\epsilon/\Lambda_{sp}} e^{-\frac{\rho z}{\Lambda_{sp}}} \left(1 - e^{-\frac{t\rho\epsilon}{\Lambda_{sp}}}\right) + \frac{P_{0,\mu}(t)}{\rho\epsilon/\Lambda_{\mu}} e^{-\frac{\rho z}{\Lambda_{\mu}}} \left(1 - e^{-\frac{t\rho\epsilon}{\Lambda_{\mu}}}\right) \quad (1)$$

This noble gas concentration will be a function of the inherited cosmogenic or non-cosmogenic noble gas concentration present at the onset of exposure to cosmic rays $C(z, 0)$, the surface production rate from spallation reactions $P_{0,sp}(t)$ and muon interactions $P_{0,\mu}(t)$, the rock density ρ , the erosion rate ϵ , and the attenuation lengths for spallation Λ_{sp} and muons Λ_{μ} . This equation assumes that the erosion rate has not varied through time.

Equation 1 highlights that, in order to determine the surface exposure age for a geologic surface from a measured cosmogenic noble gas concentration, we must know the production rate of that cosmogenic noble gas from each production mechanism. Cosmogenic nuclide production rates are a function of the elemental composition of the target mineral, element- and particle-specific nuclear reaction excitation functions (probability of a reaction as a function of energy), and the differential flux (flux as a function of energy) of cosmic ray particles (Lal & Peters, 1967; Masarik & Reedy, 1995). A common reference frame for cosmogenic nuclide production rates is the production rate at modern sea level and high latitude (SLHL), because the incoming GCR flux at high latitudes is least susceptible to geomagnetic field variations (discussed in more detail below). To determine production rates for sample sites at different locations on Earth, we apply scaling factors to the SLHL production rate for spallogenic production. These scaling factors account for variations in latitude, altitude and geomagnetic field strength. Determining accurate scaling factors for cosmogenic nuclide production is a nontrivial task because the flux of cosmic-ray particles reaching the Earth's surface varies both in space and time.

The instantaneous flux of cosmic-ray particles reaching the Earth's surface will vary spatially as a function of atmospheric depth and the configuration of Earth's magnetic field. Interactions between cosmic-ray particles and the atmosphere attenuate the cosmic-ray flux that reaches the Earth's surface (e.g., Gosse & Phillips, 2001), the greater the atmospheric depth above a particular location, the greater the attenuation of the cosmic-ray flux. Because atmospheric depth varies inversely with elevation approximately exponentially, the cosmic-ray flux and therefore cosmogenic nuclide production rates increase similarly with increasing elevation. Cosmogenic nuclide production rates approximately double with every kilometer of elevation increase (e.g., Balco et al., 2008). The atmospheric depth of a particular site can change over geologic time for two reasons. First, uplift or subsidence can change the elevation of a site and therefore its atmospheric depth. And second, changes in global climate can lead to changes in atmospheric depth, even in the absence of uplift or subsidence. There are two primary effects that changes in climate have on atmospheric depth. First, if water is redistributed between the oceans and continental ice sheets, sea level and therefore the height of the base of the atmosphere changes. Lower sea level in the geologic past will result in a lower atmospheric pressure at a given point on the Earth's surface compared to a higher sea level. Second, the atmosphere compresses when it is colder and expands when it is warmer, which

changes the atmospheric pressure at Earth's surface. A colder climate will result in a higher atmospheric pressure at a given point on the Earth's surface compared to a warmer climate. When considered together, the simultaneous effects of changes in sea level and atmospheric temperature often counteract one another: when the planet experiences a colder climate, ice sheets tend to be larger and sea level is consequently lower, and vice versa.

Position within the Earth's magnetic field matters because primary cosmic-ray particles are charged and therefore deflected by the Earth's magnetic field. The geomagnetic field is generated in the Earth's liquid outer core and extends into space, forming the magnetosphere. It is within the magnetosphere that the magnetic field interacts with charged cosmic-ray particles. We can approximate the shape of the Earth's magnetic field as a dipole field aligned with the Earth's rotation axis (also known as a geocentric axial dipole, or GAD), where geomagnetic and geographic latitudes are equivalent. Near the geomagnetic equator, vertically incident cosmic-ray particles are perpendicular to the Earth's magnetic field lines, whereas near the poles vertically-incident cosmic-ray particles are parallel to the field lines (e.g., Dunai, 2001). As a consequence, higher energies are required for cosmic-ray particles to penetrate Earth's magnetic field at lower geomagnetic latitudes (Figure 1a). Cosmogenic nuclide production rates roughly double from the equator to the poles (Masarik et al., 2001).

In addition to the spatial variations in the cosmic-ray flux reaching the Earth's surface at any point in time, this flux can also vary through time for a particular geologic surface due to variation in the Earth's magnetic field strength and external variations of incoming cosmic-ray flux. Field intensity effects on cosmogenic nuclide production rates can be parametrized using cutoff rigidity (R_c), which is the minimum momentum per unit charge needed for an incident cosmic-ray particle to penetrate the Earth's magnetic field (Dunai, 2001; Elsasser et al., 1956):

$$R_c = \frac{M_i \mu_0 c}{16\pi R^2} \cos^4(\lambda) \quad (2)$$

where M_i is the dipole moment or the strength of the Earth's magnetic field, μ_0 is the permeability of free space ($1.256 \times 10^{-6} \text{ m kg s}^{-2} \text{ A}^{-2}$), c is the speed of light ($2.99 \times 10^8 \text{ m/s}$), λ is the geomagnetic latitude assuming a GAD field (i.e., geographic latitude), and R is the radius of the Earth ($6.37 \times 10^6 \text{ m}$). Because of the shape of the Earth's dipole field, cutoff rigidities are low ($\ll 1 \text{ GV}$) at the poles and increase to a maximum at the equator ($>14 \text{ GV}$) (Figure 1). The strength of the geomagnetic field has varied throughout Earth's history. These variations are driven primarily by changes in convection in the liquid outer core which can arise on short (days to years) and long ($>>1,000$ years) timescales. During periods in Earth history of lower geomagnetic field intensity (i.e., when M_i in Equation 2 was smaller), cutoff rigidities are decreased at subpolar latitudes, and as a consequence cosmogenic nuclide production rates are greater (Figure 1). The Sun's magnetic field cycle also modulates the lower-energy portion of the galactic cosmic-ray flux (Lifton et al., 2005). However, because Solar cycles are periodic and relatively high-frequency (e.g., the 11-year solar cycle), its modulation of the galactic cosmic-ray flux can be well represented by a long-term average modulation parameter for the purposes of studying surface exposure histories with cosmogenic nuclides (Lifton et al., 2005).

In addition to secular changes in magnetic field intensity, the positions of tectonic plates have also changed on million year timescales. As a result, the cutoff rigidity and therefore production rate that applies to a particular geologic surface can change due to changes in that surface's paleolatitude independent of any changes in the geomagnetic field morphology. These changes due to tectonic plate motions will be negligible on late Quaternary timescales, but may be substantial going back deeper in time, depending on the tectonic plate in question. This is highlighted in Figure 2, where the paleolatitude of a location in India at 20°N today is demonstrated to have changed significantly over the last 70 Myr due to motion of the Indian plate. In contrast, locations in Greenland at 75°N , northern Chile at 19°S , and South Africa at 31°S today have experienced 15% or less change in paleolatitude due to the motions of the North American, South American, and African plates, respectively, over this same time period.

1.2. Existing Production Rate Scaling Schemes

Several generations of scaling schemes have been developed to determine site and sample-specific cosmogenic nuclide production rate scaling factors; in general, these scaling schemes have increased in sophistication through time (Lal, 1991, later reparametrized by Stone, 2000; Dunai, 2000, 2001; Lifton et al., 2005; Desilets & Zreda, 2003; Desilets et al., 2006; Lifton et al., 2014). Scaling models routinely used by the cosmogenic nuclide

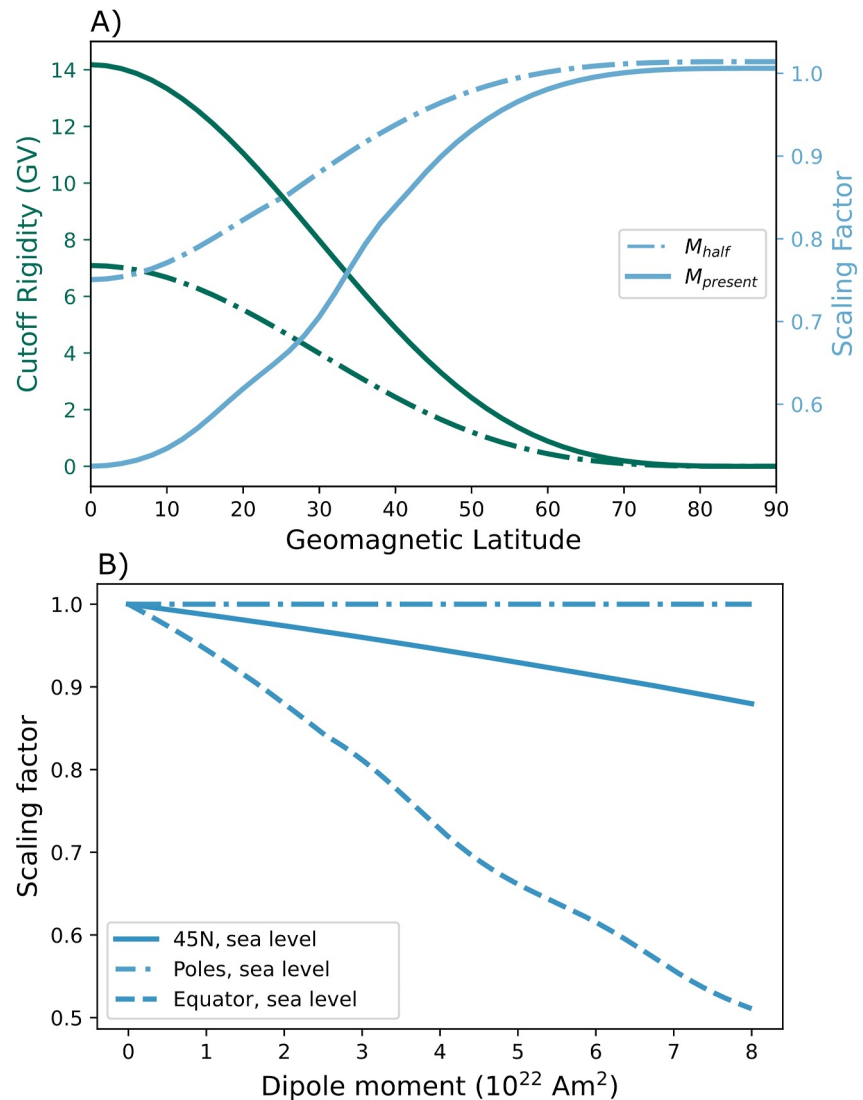


Figure 1. (a) Relationship between cutoff rigidity (R_c , dark green) and scaling factor (blue) as a function of geomagnetic latitude, calculated at sea level for ^3He in pyroxene. We show cutoff rigidities and scaling factors calculated both for the present-day magnetic field intensity (M_{present} , solid curves (IGRF-13, Alken et al., 2021)) and a field intensity that is one half that of the present day field intensity (M_{half} , dot-dashed curves). (b) Spallogenic scaling factor for cosmogenic ^3He production in pyroxene as a function of geomagnetic field intensity at different latitudes. At most latitudes, the flux of cosmic rays reaching Earth's surface decreases with increasing field intensity, decreasing cosmogenic nuclide production rates and therefore scaling factors. Scaling factors are independent of geomagnetic field intensity at the poles (dashed-dotted) because cosmic-ray particle trajectories are parallel to the field lines here and do not need to overcome the field intensity to penetrate Earth's atmosphere. Comparatively, cosmic-ray particles have a much higher energy threshold to penetrate at the equator (dashed), and exhibit a much wider range of scaling factors as a function of geomagnetic field intensity relative to the mid-latitudes (solid).

community are listed in Table 1. Lal (1991), the first published scaling scheme, quantified the effects of geomagnetic latitude and altitude on production rates predominantly using photographic emulsion data and, to a lesser extent, neutron monitor data. Stone (2000) reparameterized the Lal scaling scheme using atmospheric pressure instead of altitude ("St"), and Balco et al. (2008) applied a time-dependent magnetic field model to the Lal/Stone scaling schemes (termed "Lm" for Lal magnetic) (Table 1). Scaling schemes independent of photographic emulsion and neutron monitor data have become possible following the work of Sato and Niita (2006) and Sato et al. (2008) who developed a model referred to as PARMA (PHITS-based Analytical Radiation Model in the Atmosphere, where PHITS stands for Particle and Heavy Ion Transport code System) to calculate the flux of

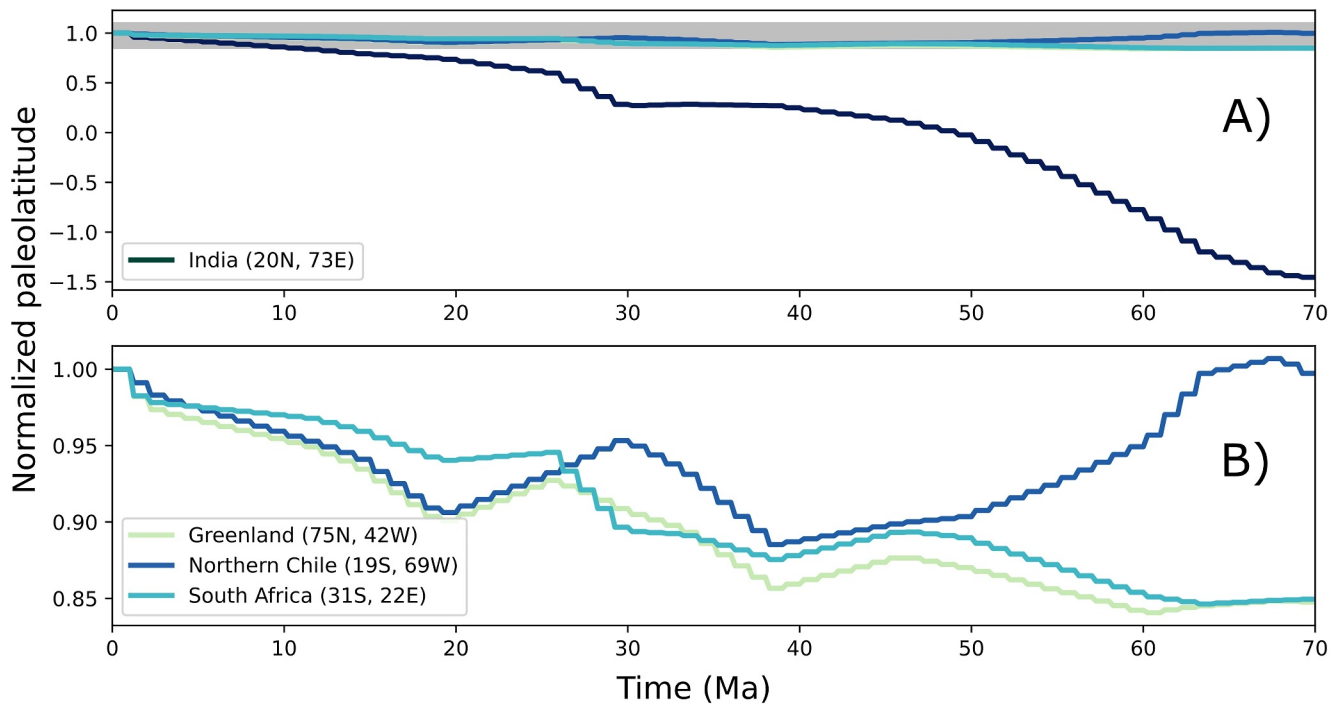


Figure 2. (a) Paleolatitude changes over 70 Myr for four locations on different tectonic plates and at different modern day latitudes, normalized to the present day latitude: India (20°N, dark blue), Greenland (75°N, light green), northern Chile (19°S, blue), and South Africa (31°S, teal). The gray box denotes the normalized paleolatitude range shown in (b). Paleolatitude data in 1 Myr bins are calculated with the PmagPy apwp program (Tauxe et al., 2016), which uses the reconstruction of Besse and Courtillot (2002). (b) Normalized paleolatitudes for locations in Greenland, northern Chile, and South Africa, shown over a smaller range of normalized paleolatitudes than in (a).

neutrons and protons at a given location and point in time. These analytical functions fit modeled energy spectra of cosmic-ray particles from 1 MeV to 200 GeV to estimate the atmospheric cosmic-ray particle flux. Using these models of atmospheric cosmic-ray flux and the elemental reaction excitation functions of Reedy (2013), Lifton et al. (2014) developed a scaling scheme referred to as the “LSDn” model. In addition to the updates to cosmic-ray flux estimation, LSDn also incorporated newer, higher resolution geomagnetic models for the last 2 Ma (Korte et al., 2009; Laj et al., 2004; Ziegler et al., 2011). This enables the LSDn scaling scheme (when paired with geologically calibrated production rates) to be used for calculating scaling factors of cosmogenic nuclides with time-varying geomagnetic field models from the present day to 2 Ma. In similar fashion to Lm, LSDn uses an average geomagnetic field strength for time periods prior to 2 Ma.

1.3. Geomagnetic Field Models

In order to calculate cutoff rigidities (R_c , Equation 2) for cosmogenic nuclide scaling models, we need to know both the direction and strength of the geomagnetic field. The field direction is commonly described using two components: declination, the angle between magnetic north and geographic north, and inclination, the dip angle between the horizontal plane and the magnetic field vector. Paleomagnetic records of Earth's ancient magnetic field in artifacts, rocks, and sediments can be used to constrain changes in paleodirection and paleointensity. For the purpose of estimating cosmogenic nuclide production rates, there are broadly two types of time-dependent geomagnetic field models available: global field models using spherical harmonic and b-splines, and geocentric dipole moment time-series models.

For the latest Quaternary, detailed global, time-dependent geomagnetic field models (i.e., spherical harmonic models) are available. These models are essentially a suite of time-dependent spherical harmonic terms, which means they can be used to solve for each component of the magnetic field. For example, the original LSDn scaling framework (Lifton et al., 2014) uses two such models, CALS3k.3 that spans the past 3,000 years (Korte et al., 2009) and CALS7k.2 that spans the last 7,000 years (Korte & Constable, 2005), which are built using a combination of archaeomagnetic, lava, and sedimentary paleomagnetic data. Extending spherical harmonic

Table 1

Summary of Commonly Used, Paleomagnetically Corrected Production Rate Scaling Model Frameworks as Presently Implemented in Version 3 of the Production Rate Calculator Originally Described in Balco et al. (2008), Compared to the Implementation of the LSDn Scaling Model as Implemented in the SPRITE Framework Presented Here

Model	Time period	Magnetic field model type ^a	Magnetic field model source	R _c method	Latitude used for R _c	Atmospheric pressure conversion
Lm	2010 CE–1950 CE	SH	International Geomagnetic Reference Model (IGRF) of (1)	Eq. 1 of (6, 7) and Eq. 2 of (8)	Geomagnetic	Eq. 1 of (10)
	1950 CE–14 ka	SH	SHA.DIF.14k of (2)	Eq. 1 of (6, 7) and Eq. 2 of (8)	Geomagnetic	Eq. 1 of (10)
	14 ka–2 Ma	GAD	GLOPIS-75 of (3); PADM2M of (4)	Eq. 1 of (6, 7) and Eq. 2 of (8)	Geographic	Eq. 1 of (10)
	>2 Ma	GAD _{constant}	GLOPIS-75 of (3); PADM2M of (4)	Eq. 1 of (6, 7) and Eq. 2 of (8)	Geographic	Eq. 1 of (10)
LSDn	2010 CE–1950 CE	SH	International Geomagnetic Reference Model (IGRF) of (1)	Trajectory tracing of (6)	Geomagnetic	Eq. 1 of (10) or ERA40 (11)
	1950 CE–14 ka	SH	SHA.DIF.14k of (2)	Trajectory tracing of (6)	Geomagnetic	Eq. 1 of (10) or ERA40 (11)
	14 ka–2 Ma	GAD	GLOPIS-75 of (3); PADM2M of (4)	Eq. 1 of (6, 7) and Eq. 2 of (8)	Geographic	Eq. 1 of (10) or ERA40 (11)
	>2 Ma	GAD _{constant}	GLOPIS-75 of (3); PADM2M of (4)	Eq. 1 of (6, 7) and Eq. 2 of (8)	Geographic	Eq. 1 of (10) or ERA40 (11)
LSDn in SPRITE framework	2010 CE–1950 CE	SH	International Geomagnetic Reference Model (IGRF) of (1)	Eq. 1 of (6, 7) and Eq. 2 of (8)	Paleolatitude of (9)	Eq. 1 of (10), ERA40 (11), or (12)
	1950 CE–14 ka	SH	SHA.DIF.14k of (2)	Eq. 1 of (6, 7) and Eq. 2 of (8)	Paleolatitude of (9)	Eq. 1 of (10), ERA40 (11), or (12)
	14 ka–2 Ma	GAD	GLOPIS-75 of (3); PADM2M of (4)	Eq. 1 of (6, 7) and Eq. 2 of (8)	Paleolatitude of (9)	Eq. 1 of (10), ERA40 (11), or (12)
	2 Ma–70 Ma	GAD	MCADAM1.b_cosmo, based on (5)	Eq. 1 of (6, 7) and Eq. 2 of (8)	Paleolatitude of (9)	Eq. 1 of (10), ERA40 (11), or (12)

Note. St = Lal (1991), Stone (2000), Nishiizumi et al. (1989), modified by Balco et al. (2008); LSDn = Lifton et al. (2014), Lifton (2016). (1) IGRF-11 (Finlay et al., 2010); (2) Pavón-Carrasco et al. (2014); (3) Laj et al. (2004); (4) Ziegler et al. (2011); (5) Bono et al. (2022); (6) Elissasser et al. (1956); (7) Dunai (2001); (8) Lifton et al. (2008); (9) Besse and Courtillot (2002); (10) Stone (2000); (11) Uppala et al. (2005); (12) Valdes et al. (2021). ^aSH: spherical harmonic field model. GAD: geocentric axial dipole approximation of field model. GAD_{constant}: a GAD field with mean dipole moment for the given time period.

models further back into the Quaternary and beyond is challenging due to an increasing sparsity of high quality, high resolution, full-vector paleomagnetic records with increasing age. Dipole moment time-series models simplify the description of the magnetic field to a geocentric axial dipole which varies over time. These models range in coverage from tens of thousands of years (e.g., 0–75 ka with GLOPIS-75, derived from sedimentary paleomagnetic data; Laj et al., 2004) to several million years (e.g., 0–2 Ma with PADM2M, derived from sedimentary paleomagnetic data; Ziegler et al., 2011) or billions of years (e.g., 50ka–4.2 Ga with MCADAM, derived from igneous and metamorphic paleomagnetic data; Bono et al., 2022).

Some assumptions must be made in order to apply geomagnetic models that cover older time periods to estimate cosmogenic nuclide production rates. For instance, the LSDn scaling scheme (Lifton et al., 2014) extends beyond 7 ka by assuming a geocentric axial dipole (GAD) field. The GAD model is a first-order approximation of the geomagnetic field which describes a magnetic field generated by a single magnetic dipole at the center of the Earth that is aligned with its rotation axis (Butler, 1992). While the morphology of the geomagnetic field varies throughout time, it has been illustrated that GAD is a robust approximation of the average field over timescales longer than ~10 Kyr (e.g., McElhinny, 2007) that appears to also be robust for most of Earth's history (Biggin et al., 2020). Assuming a GAD field, Lifton et al. (2014) trace the trajectories of anti-protons and, using knowledge of relative variation in paleointensity estimated from sedimentary records, estimate the amount of incoming cosmic-ray protons at different magnetic latitudes. Finally, Lifton et al. (2014) use a sixth-order polynomial of latitudinal cosines and the hybrid paleointensity reconstructions GLOPIS-75 and PADM2M (spanning the past 75 Kyr and 2 Myr, respectively) to extend the model beyond 7 ka.

Extending geomagnetic models further back than the Quaternary requires the use of absolute paleointensity recorded in igneous and metamorphic rocks as records of relative paleointensity from sediment cores become sparse. One such model, MCADAM (Bono et al., 2022), is a continuous geomagnetic model based on data from the PINT v8.1.1 database (Bono et al., 2021), which contains 4,648 absolute paleointensity records spanning 50 ka to 4.2 Ga. MCADAM uses a combination of non-parametric and Monte Carlo resampling, along with time-adaptive smoothing kernels that adjusts the minimum amount of time needed for a time average, to yield an estimate of the strength of the dipole field. The MCADAM models of Bono et al. (2022) define a minimum averaging kernel of 50 Kyr, which extends during intervals with sparse data coverage.

Paleomagnetic data from the Quaternary benefit from sufficient coverage in time and space to define global spherical harmonic models that capture secular variation, which is variation in the magnetic field on timescales <10 Kyr. Due to the relaxation, alteration or presence of spurious magnetic carriers, paleomagnetic data become increasingly sparse further back in time, precluding the definition of time-varying spherical harmonic models. In the case of PADM2M, absolute paleointensity data are supplemented with relative paleointensity stacks in order to constrain secular variation; in MCADAM, only the average dipole is estimated, with the uncertainty representing a combination of secular variation and measurement variability (Bono et al., 2022; Ziegler et al., 2011).

2. Methods

Here we describe the model framework Scaling Production Rates In deep Time, or SPRITE, which can be used to calculate cosmogenic noble gas production rate scaling factors for the last 70 Ma. SPRITE is built upon the LSDn scaling model for spallogenic production, the Balco (2017) model for muogenic ^{21}Ne production, and incorporates pre-Quaternary changes in geomagnetic field strength, paleolatitude, and paleoatmospheric depth. Figure 3 demonstrates the functionality of SPRITE. The default model settings enable scaling factors to be calculated back to 70 Ma, primarily because we developed SPRITE to determine production rates during the eruption of the Deccan Traps across the Cretaceous-Paleogene boundary. In principle, the code can be modified to extend scaling factor calculations that account for paleointensities, paleoatmospheric depths, and paleolatitudes to 200 Ma (i.e., the full extent of the paleolatitude reconstruction we use), for paleointensities and paleoatmospheric depths to 540 Ma (i.e., the full extent of the climate simulations we use), or only for time-varying magnetic field strengths to ~4.2 Ga (i.e., the full duration of MCADAM). However, the latter two extensions are not recommended due to data sparsity and increased uncertainty throughout the Precambrian paleointensity record (Bono et al., 2021).

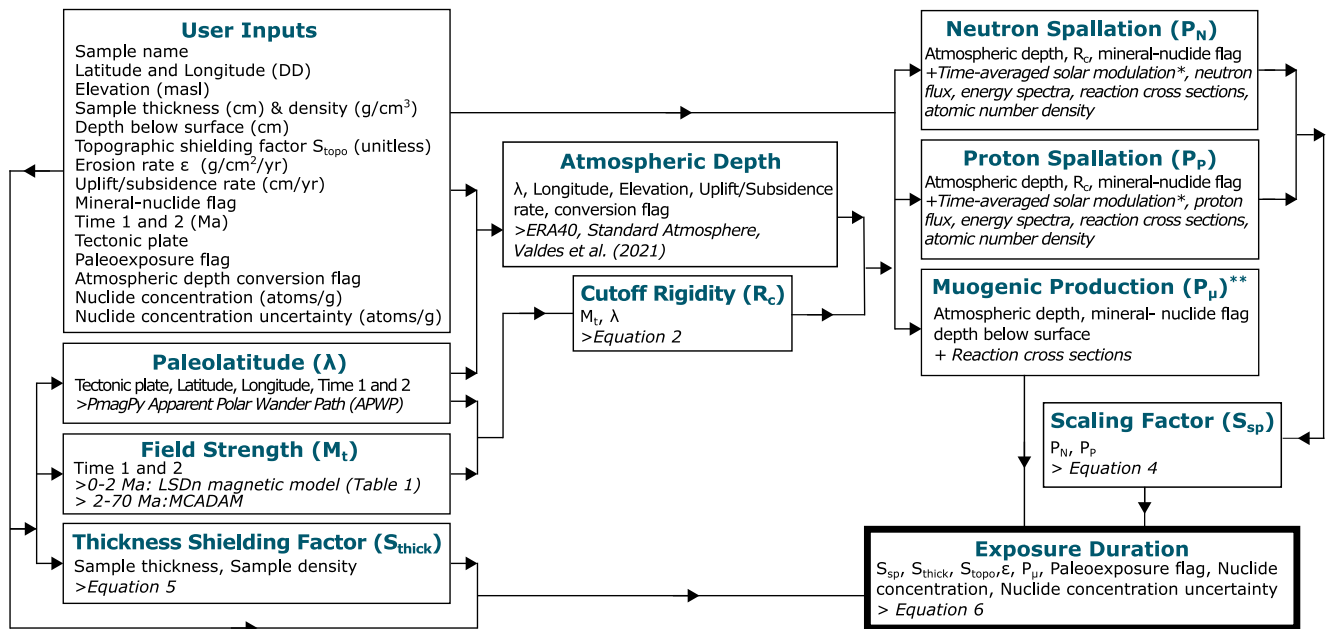


Figure 3. SPRITE workflow diagram, demonstrating how user inputs as well as information from various model frameworks interact and lead to an exposure duration calculation. Following the user input box, each box in the diagram indicates (1) the parameter being calculated in bold teal font, (2) the parameters derived from the preceding boxes that are necessary for that calculation in black regular font, and (3) the additional models and/or information within SPRITE that are used to perform that calculation in black italic font.

2.1. User Inputs

In order to calculate production rate scaling factors as well as exposure ages, SPRITE requires several sample specific pieces of information that are summarized in the “User Inputs” box in Figure 3. This model setup information is largely the same as what other existing exposure age calculators require (e.g., Balco et al., 2008; Marrero et al., 2016; Martin et al., 2017), and their use in calculations is described in greater detail below. There are, however, several model inputs that are unique to SPRITE that are worth describing in more detail here. First, because SPRITE accounts for a sample's paleolatitude using a plate tectonics reconstruction, the tectonic plate that sample is from must be specified. Second, for the pre-Quaternary applications for which we envision SPRITE being used, a sample's exposure to cosmic-ray particles can either be continuous through time until the present day, or it can be a paleoexposure, meaning that the geologic surface of interest was exposed to cosmic-ray particles for some duration of time in the geologic past and subsequently buried, terminating cosmogenic nuclide production. The SPRITE framework requires the type of cosmogenic nuclide exposure history to be specified (continuous exposure or paleoexposure), as well as when in the geologic past exposure to cosmic-ray particles began, in the case of paleoexposures (“Time 1”, Figure 3). The user can also impose an upper limit on the duration of exposure (“Time 2,” Figure 3) if they have geologic evidence to support such a limit, but this is not required and primarily only functions to increase the computational efficiency of the model calculations. Providing a relevant duration of time over which to run the model is consistent with how the original LSDn scaling scheme computation works, because prior to an exposure age calculation, it is not a priori known how many time-dependent scaling factors are needed. In the SPRITE framework, once the exposure age calculation begins, the integration will only use as many bins (i.e., time-dependent scaling factors) as needed to achieve the observed nuclide concentration. The requirements for all model inputs are described in detail in the documentation provided with the SPRITE code repository as well as an example input spreadsheet.

2.2. Cutoff Rigidity Calculation

Cutoff rigidity (Equation 2) is one of the first parameters that SPRITE calculates in order to determine production rate scaling factors (Figure 3). To calculate cutoff rigidities, SPRITE first determines the time-varying geomagnetic dipole strength and paleolatitude for each sample. From 0 to 2 Ma, we use the same magnetic field models as LSDn (Table 1). Beyond 2 Ma, the dipole strength is obtained using a modified continuous

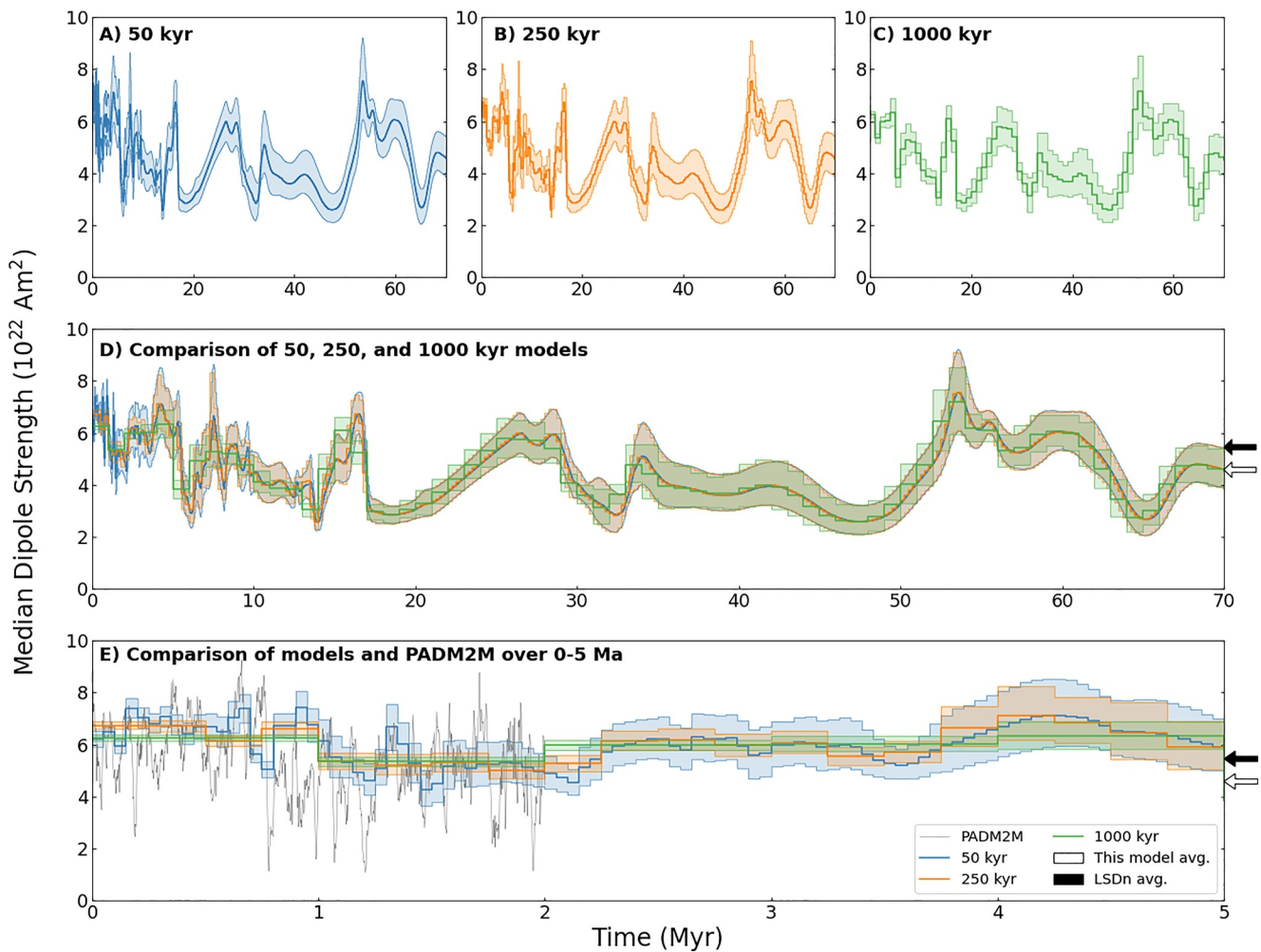


Figure 4. Continuous geomagnetic dipole strength models with resolutions of (a) 50 Kyr, (b) 250 Kyr, and (c) 1,000 Kyr. (d) Comparison of the models with different temporal resolutions shown in (a)–(c). (e) The same dipole models as (d) from 5 Ma to present, during which there are the most significant deviations between the 50 and 250 Kyr temporal resolution models. PADM2M, one of the models used in LSDn, is shown in gray. Arrows in (d) and (e) show the long-term average of MCADAM (present to 70 Ma in (d), $4.6 \times 10^{22} \text{ Am}^2$; and present to 5 Ma in (e), $6.06 \times 10^{22} \text{ Am}^2$ —white arrow) compared to the average used in the LSDn scaling scheme for exposures prior to 2 Ma ($5.5 \times 10^{22} \text{ Am}^2$ —black arrow). The uncertainty shown by the shaded regions in the models represent the interquartile prediction bounds of MCADAM. SPRITE uses the same magnetic model as LSDn from 0 to 2 Ma (including PADM2M, gray line), and MCADAM (250 Kyr resolution, orange line) from 2 to 70 Ma. The magnetic field model implemented from 0 to 14 ka (Table 1) in LSDn and the SPRITE framework is not shown here as it would not be discernible at this scale.

geomagnetic dipole strength model that covers 2–70 Ma, following the methods of Bono et al. (2022). We elected to base our models on MCADAM1.b ($N_{\text{site}} = 2,327$) which only includes data that meet at least three of the Quality of Paleointensity (Q_{PI}) criteria (Biggin & Paterson, 2014) as it has been shown that filtering the data in this way balances data quality with data availability (Bono et al., 2021, following the analysis of Biggin et al. (2015)). In addition, the model excludes data from any sites that are described as having transitional polarity. We use the median rather than the mean paleointensity because paleointensities do not follow a normal distribution.

The MCADAM approach was modified to set the minimum temporal averaging kernel to match the sampling step size, creating synthetic “bins” for time averaging. These dimensions are kernel minimums since the MCADAM approach expands averaging kernels until a sufficient quantity of data is averaged. We tested MCADAM model time bin sizes of 50 Kyr (matching the temporal resolution of Bono et al., 2022), 250 Kyr, and 1,000 Kyr. Bin sizes of 1,000 Kyr resulted in coarser geomagnetic models while the 50 and 250 Kyr gave nearly indistinguishable results (Figure 4). Moreover, we find scaling factors calculated using 250 Kyr bins are consistent with scaling

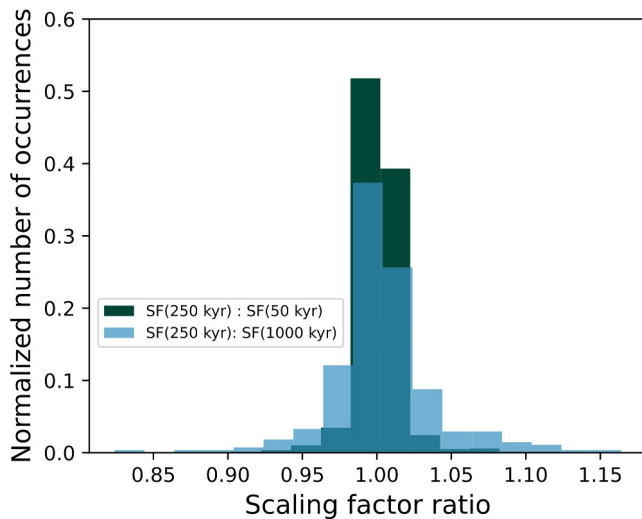


Figure 5. Comparison of scaling factors calculated using different temporal resolutions for MCADAM1.b_cosmo within SPRITE, calculated for cosmogenic ^3He production in pyroxene at 23°N and sea level. Histograms show the ratios of scaling factors over the model time domain of SPRITE (2–70 Ma) comparing the default model resolution of 250–50 Kyr bin sizes (dark green) and 1,000 Kyr bin sizes (light blue). A ratio of 1 indicates that scaling factors from two bin sizes are equal. Differences in scaling factor between the 250 and 50 Kyr resolution models are generally smaller than the differences in scaling factor between the 250 and 1,000 Kyr resolution models.

factors calculated using the 50 Kyr bin size, while using 1,000 Kyr bins results in higher scatter around the full resolution model (Figure 5). For computational purposes, we use a time bin size of 250 Kyr from 2 to 70 Ma as the default in SPRITE (hereafter termed MCADAM1.b_cosmo), but note that the user can change the MCADAM resolution in SPRITE if desired. From 0 to 2 Ma, we use the same temporal resolution as implemented in LSDn, which is variable and ranges between 10 and 2000 years.

To obtain a sample's paleolatitude, SPRITE calls PmagPy (Tauxe et al., 2016), an open-source Python toolkit that enables the analysis of paleomagnetic and rock magnetic data archived in the Magnetism Information Consortium (MagIC) database (<https://www2.earthref.org/MagIC>). SPRITE uses the present-day latitude, longitude, and tectonic plate of a cosmogenic sample site as inputs to the apparent polar wander path (apwp) program in PmagPy, which uses the APWP kinematic model of Besse and Courtillot (2002) for major tectonic plates (Africa, Antarctica, Australia, Eurasia, Greenland, India, North America, and South America) over the last 200 Myr. The reconstruction presented in Besse and Courtillot (2002) is a simple plate model and although other more complex reconstructions exist (e.g., Torsvik et al., 2012), the differences between the models in terms of the effect on cosmogenic production rate is minor. We do note, however, that the Besse and Courtillot (2002) model cannot reconstruct paleolatitudes for sites on the Pacific Plate. When using SPRITE for sites located on the Pacific Plate, modern latitude should be used. SPRITE uses geomagnetic dipole strength (M_t) and paleolatitude (λ) as described above to calculate time-varying R_c values for each sample using Equation 2 (Figure 3).

2.3. Atmospheric Depth Calculation

Alongside cutoff rigidity, atmospheric depth is also one of the first parameters that SPRITE calculates to determine production rate scaling factors (Figure 3). When a sample is collected for cosmogenic nuclide analysis, information about sample elevation must be converted into atmospheric pressure to account for atmospheric attenuation on production rates. SPRITE converts elevation to atmospheric pressure using one of three approaches. One conversion in SPRITE applies the U.S. Standard Atmosphere 1D Model (National Oceanic and Atmospheric Administration, 1976):

$$P(z) = P_s \exp \left\{ -\frac{gM}{Re} (\ln(T_s) - \ln(T_s - \epsilon z)) \right\} \quad (3)$$

where $P(z)$ is pressure (hPa) as a function of depth z (masl), P_s is sea level pressure (hPa), g is the acceleration due to gravity (m/s^2), M is the molar weight of air (g/mol), R is the gas constant (J/K/mol), ϵ is the adiabatic lapse rate (K/m), and T_s is sea level temperature (K). This equation is meant to approximate mean annual atmospheric conditions at midlatitudes. However, atmospheric reanalysis data sets demonstrate that pressure deviations from the standard atmosphere model exist that can lead to percent level differences in cosmogenic nuclide production rates at sea level, and systematically underestimate production rates at high altitudes (Balco et al., 2008; Stone, 2000). As a consequence, existing scaling models often apply corrections to Equation 3 using spatial variations in mean sea level pressure and temperature captured in reanalysis data sets. For example, LSDn (Lifton et al., 2014) utilizes European Center for Medium-Range Weather Forecasts (ECMWF) 40-year reanalysis (ERA-40) data (Uppala et al., 2005). We therefore also include a conversion option in SPRITE that utilizes the ERA-40 data set to correct Equation 3.

The ERA40 data set reports atmospheric pressure and temperature conditions between 1957 and 2002. Over geologic timescales, however, we expect the atmospheric pressure and temperature structure to vary because of changes in global climate. The effects of past climate conditions on cosmogenic nuclide production rates were investigated by Staiger et al. (2007) for the last glacial maximum (LGM), representative of some of the coldest

temperatures and lowest sea levels in the last 70 Ma. Using an atmosphere-only general circulation model (GCM), Staiger et al. (2007) found that significant differences in production rates at the LGM relative to today were only expected at high elevations (>4 km) and primarily driven by changes in atmospheric temperature rather than sea level, as well as near ice sheet margins due to a combination of atmospheric cooling and katabatic winds. Similarly, Cuzzone et al. (2016) used coupled atmosphere-ocean GCM simulations at 3 ka intervals from 21 ka to the present to investigate the effects of changing atmospheric thickness since the LGM on production rates in Scandinavia, and found that the effect of changing climate conditions with respect to their exposure ages was negligible relative to other processes impacting production rates at their study sites (namely, post-glacial rebound). For times before the LGM, the temporal resolution with which GCM simulations are available decreases significantly.

For the SPRITE framework, we use the simulations of Valdes et al. (2021) to estimate the effects of changing climate on atmospheric pressures and temperatures and therefore atmospheric depths over geologic timescales. Valdes et al. (2021) carried out simulations over the entire Phanerozoic with the atmosphere-ocean-vegetation coupled model HADCM3, utilizing the atmospheric CO₂ reconstruction of Foster et al. (2017) and the paleogeographic reconstructions of Scotese (2016). Over the last 70 Ma (i.e., the model timespan of SPRITE), the Valdes et al. (2021) simulations are carried out in ~5 Ma intervals; to our knowledge, this is the highest temporal resolution of GCM simulations available from a single model over this time period. We then follow a similar approach as described above for the reanalysis data sets, using the time- and spatially varying sea level pressures and temperatures from the relevant 5 Ma timestep in the Valdes et al. (2021) model simulations to correct Equation 3.

In addition to atmospheric changes that can cause site-specific atmospheric pressures to vary through time, we must also consider changes in the elevation of sample sites through time. For the longer geologic timescales that SPRITE is applicable to, virtually all landscapes will experience uplift and/or subsidence, either throughout their exposure duration in the case of long-duration continuous exposures (Dunai et al., 2005; Ritter, Stuart, et al., 2018), or since their exposure time in the case of paleoexposures. Both continuous and paleoexposures require an estimate of the elevation when the surface was first exposed, and for continuous exposures an assumption of the rate at which uplift or subsidence occurred. Because production rates increase approximately 1% for every 10 m increase in elevation (Lal, 1991), uncertainties in site elevation can lead to significant uncertainties in production rate. We acknowledge that this is the most difficult parameter to account for when extending cosmogenic nuclide applications to longer, pre-Quaternary timescales. Indeed, obtaining accurate and precise paleoelevation information remains one of the grand challenges in geologic research more generally (e.g., Huntington & Klepeis, 2018; Mulch, 2016). Existing applications of cosmogenic nuclides on these timescales have used other regional or local geologic observations to attempt to estimate paleoelevations and/or rates of elevation change and their effects on production rates (e.g., Blard et al., 2005, 2019; Dunai et al., 2005; Libarkin et al., 2002). In the absence of larger scale paleoelevation models or data sets that can be incorporated into SPRITE (e.g., comparable to MCADAM), we enable the user to provide an uplift or subsidence rate (cm/yr) should they have independent constraints on what such rates might be, in the case of a continuous exposure history. The model will then re-calculate the site elevation at each time step before calculating scaling factors. For comparatively shorter paleoexposures, the user can also test the impact of different potential paleoelevations during exposure by changing the sample elevation in SPRITE.

2.4. Theoretical Production From Spallation and Muons

With atmospheric depth and time-varying cutoff rigidities calculated, SPRITE calculates site-specific, time-varying theoretical spallogenic and muogenic production rates (Figure 3). The SPRITE framework calculates the theoretical spallation production rates of ³He and ²¹Ne from neutron and proton bombardment using the LSDn scaling scheme (Lifton et al., 2014). We summarize how the LSDn scaling scheme calculates theoretical spallation production rates here. In order to quantify production rates from spallation reactions, we need to know both the cosmic-ray particle flux and reaction excitation functions. To calculate the flux of cosmic-ray neutrons and protons, LSDn uses the PARMA analytical functions of Sato et al. (2008). These analytical functions fit modeled energy spectra of cosmic-ray particles (i.e., 1 MeV–200 GeV, the range of energies possible for incoming cosmic-ray particles) to estimate the cosmic-ray flux through the lower 20 km of the atmosphere (Sato et al., 2008). The estimated neutron and proton flux for each sample's location in each time step is adjusted using the previously calculated cutoff rigidities (Section 2.2) and atmospheric depths (Section 2.3). Reaction excitation functions are

the probability as a function of energy that a nuclear reaction, such as spallation via neutrons and protons, will occur on some target element (e.g., Si, O, Mg) and produce a cosmogenic nuclide like ^3He and ^{21}Ne . SPRITE uses the compilation of the experimentally determined neutron and proton reaction excitation functions presented in Reedy (2013), which include excitation functions for ^3He and ^{21}Ne production rate calculations. Because the abundances of target elements varies between different minerals, LSDn also specifies the number density of target atoms in different minerals to calculate the mineral- and nuclide-specific production rate. SPRITE uses the differential particle flux, excitation functions, and number densities to calculate a theoretical production rate from neutrons P_n and a theoretical production rate from protons P_p for each sample and for each time step.

Muogenic production of ^3He and ^{21}Ne are calculated differently from one another in the SPRITE framework. Until recently, we did not have unambiguous observations demonstrating that cosmogenic ^3He production by muons occurs (Farley et al., 2006). As a consequence, it has often been assumed in existing scaling models that there is no ^3He production by muons (Balco et al., 2008). However, Larsen et al. (2021) observed a clear signature of muogenic ^3He in pyroxene and ilmenite from a 300 m deep drill core of the Columbia River Basalt. In contrast to other cosmogenic nuclides, Larsen et al. (2021) found that the muogenic contribution to surface ^3He production is small ($\sim 0.1\text{--}0.2\%$, compared to $\geq \sim 2\%$ for some other nuclides), and that the calculated attenuation length for muogenic ^3He production, $8,780\text{ g/cm}^2$, is a factor of ~ 2 larger than for other nuclides. In the absence of other empirical or modeled constraints on the muogenic production of ^3He , SPRITE incorporates the empirical parameter estimates for muogenic ^3He production in pyroxene from Larsen et al. (2021).

In contrast to ^3He , muogenic production of ^{21}Ne is well characterized and can account for $< \sim 2\%$ of ^{21}Ne surface production (Balco et al., 2019). We follow the exponential approximation from Balco (2017), using ^{21}Ne reaction cross sections (Balco et al., 2019; Fernandez-Mosquera et al., 2008) and varying muon attenuation lengths as a function of atmospheric pressure (Balco, 2017). Balco (2017) demonstrated that for surface exposure dating applications, this approximation is sufficiently accurate, and it is computationally inexpensive relative to a full numerical integration to calculate subsurface muon fluxes (Heisinger et al., 2002a, 2002b). We also only calculate fast muon production of ^{21}Ne because Balco et al. (2019) showed production via negative muon capture is insignificant. For erosion rate calculations, Balco (2017) concluded that muon attenuation lengths for ^{10}Be and ^{26}Al are site specific and vary based on erosion rate. However, the uncertainties in the geological assumptions made when estimating erosion rates over deep timescales using stable nuclides, which take much longer to reach equilibrium between production and erosion relative to radionuclides with short half lives, are much higher than the actual numerical method used to calculate erosion rates (Balco, 2017). Therefore, we implement the same muogenic production model as Balco et al. (2019) for erosion rate calculations, which only varies attenuation length as a function of atmospheric pressure, as this method captures the magnitude of muogenic contribution accurately enough within the uncertainties of the geological context of erosion rate studies.

2.5. Scaling Factor Calculations

With the theoretical production rates from neutrons (P_N) and protons (P_P) calculated for each time bin for each sample, we can calculate a dimensionless spallation scaling factor, S_{sp} , by dividing the sum of these theoretical production rates by a theoretical reference spallogenic production rate P_R calculated for SLHL conditions (Equation 4, Lifton et al., 2014). Because R_c is essentially independent of magnetic field strength at the poles, P_R is not time-varying and reflects the maximum spallogenic production rate possible at sea level.

$$S_{sp} = \frac{P_N + P_P}{P_R} \quad (4)$$

A separate thickness scaling factor is calculated after Balco et al. (2008) to correct for the attenuation of spallogenic production due to the thickness of the sample:

$$S_{thick} = \frac{\Lambda_{sp}}{\rho z_{max}} \left(1 - \exp\left(\frac{-\rho z_{max}}{\Lambda_{sp}}\right) \right) \quad (5)$$

where Λ_{sp} is the spallation attenuation length, ρ is the sample density, and z_{max} is the sample thickness. This is inherently different from sample shielding due to burial after exposure in an eroding surface, which is discussed below.

2.6. Exposure Age Calculations

While the primary purpose of SPRITE is to calculate time-varying production rate scaling factors for pre-Quaternary cosmogenic nuclide applications, SPRITE can also calculate an exposure duration using the time-dependent scaling factors. To calculate an exposure duration given some finite erosion rate, SPRITE will solve Equation 6 modified after Balco et al. (2008):

$$N = S_{thick} S_{topo} P_{ref,cal} \int_0^T S_{sp}(t) \exp\left(\frac{-\epsilon t}{\Lambda_{sp}}\right) dt + \int_0^T P_{\mu} \exp\left(\frac{-\epsilon t - z/2}{\Lambda_{\mu}}\right) dt \quad (6)$$

where S_{thick} is the shielding correction due to thickness of sample (Equation 5), S_{topo} is the shielding correction due surrounding topography, $P_{ref,cal}$ is the geologically calibrated SLHL reference production rate (discussed below), S_{sp} is the spallation scaling factor at time t for a bin size of dt (Equation 4), ϵ is the user-specified erosion rate in $\text{g}/\text{cm}^2/\text{yr}$, and Λ_{sp} is the effective spallation attenuation length. The calibrated reference production rate $P_{ref,cal}$ in Equation 6 is distinct from P_R in Equation 4, which is a theoretically calculated SLHL production rate. For ^3He production in olivine/pyroxene and ^{21}Ne production in quartz, we use the primary calibration production rate data set summarized by Borchers et al. (2016) and the production rate calibration data set of Fenton et al. (2019), respectively, to recalculate $P_{ref,cal}$ using LSDn. All the calibration sites for ^3He in olivine/pyroxene and ^{21}Ne in quartz are younger than 250 ka (Borchers et al., 2016, and references therein; Fenton et al., 2019). Over this time period, the magnetic models in LSDn and the SPRITE framework are the same; therefore using the $P_{ref,cal}$ calculated with LSDn in the SPRITE framework is appropriate and internally consistent. The second term in Equation 6 refers to muogenic production P_{μ} , which is scaled for ^{21}Ne in quartz after Balco (2017), as described above. We also include a nominal (i.e., not calculated) value of P_{μ} and Λ_{μ} to calculate muogenic ^3He production based on the empirical observations of Larsen et al. (2021).

If the user has specified that their samples experienced continuous exposures, SPRITE operates analogously to other production rate and exposure age calculators by beginning calculations with Equation 6 on the present day and continuing to add increments of cosmogenic nuclide production further into the past until the predicted and observed cosmogenic nuclide concentrations are equal. In the case of paleoexposures, SPRITE begins calculations with Equation 6 at the user-specified time in the geologic past and adds increments of cosmogenic nuclide production toward the present day until the predicted and observed cosmogenic nuclide concentrations are equal.

2.7. Example Applications

We examine the efficacy of SPRITE with existing cosmogenic noble gas data sets characterized by pre-Quaternary exposures from two different geologic settings. The first data sets, which come from Dunai et al. (2005) and Evenstar et al. (2017), consist of cosmogenic ^{21}Ne measurements in quartz and ^3He measurements in pyroxene from boulders exposed in a series of alluvial surfaces that collectively make up the Atacama paleosurface in Chile, also termed the Pacific Paleosurfaces (PPS; Cooper et al., 2016; Dunai et al., 2005; Evenstar et al., 2009, 2017; Hoke et al., 2007; Ritter, Binnie, et al., 2018; Ritter, Stuart, et al., 2018). The apparent exposure ages of boulders from these surfaces reported in the original publications range from 0.05 to ~36 Ma, and are inferred to represent the diachronous surface formation from the Oligocene to the Pleistocene through a combination of aggradational and degradational processes. We assume the authors' original interpretations of simple exposure (no burial, no erosion) and negligible inheritance are correct. We carry out two different exposure age comparisons using these data. One exposure age calculation uses the present-day sample elevations from these data sets as their constant elevation through time, while the second calculation uses the uplift rate applied by Dunai et al. (2005) of 0.004 cm/yr . By carrying out the former, constant elevation calculation, it is not our intention to change or question the original interpretations of Dunai et al. (2005) and Evenstar et al. (2017); rather, we do this to isolate the effects of the time-varying models (e.g., paleointensity, paleolatitude, paleo-atmospheric depth) built into the SPRITE framework and evaluate the magnitude of their effects on exposure ages for long duration (many Ma) surface exposures.

The second data set from Libarkin et al. (2002) consists of a ^{21}Ne depth profile that records the Oligocene paleoexposure history of Fish Canyon Tuff in Colorado, USA, before it was buried by the Carpenter Ridge Tuff ~500 ka later. The observed cosmogenic ^{21}Ne concentrations are much lower than would be expected for this

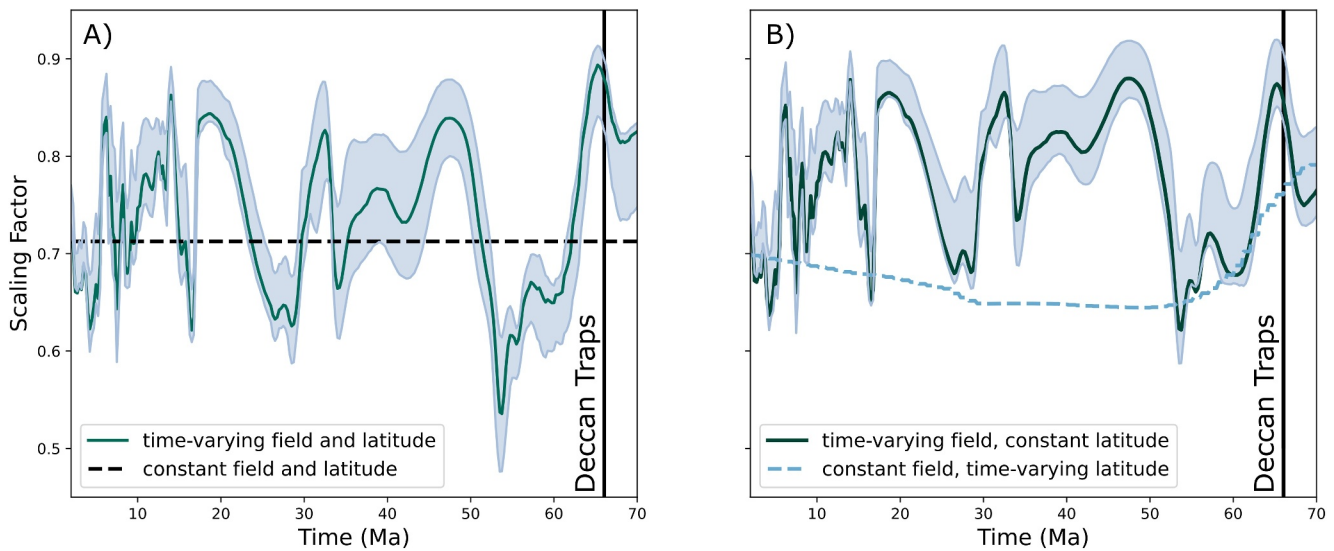


Figure 6. Spallogenic scaling factors for cosmogenic ^3He production in pyroxene for a present-day location in India (23°N , 70°E) at sea level. (a) Scaling factors from 2 to 70 Ma calculated using SPRITE (solid green) compared to the scaling factor calculated assuming the average field strength used by the LSDn scaling scheme for exposure prior to 2 Ma (dashed black). (b) Scaling factors through time calculated using SPRITE and assuming either a time-varying magnetic field with constant latitude (solid dark green), or assuming the long term average paleointensity derived from MCADAM and time-varying latitude (dashed light blue). The time period over which the Deccan Traps erupted is represented by the black vertical bar. Light blue bands represent 25% and 75% prediction bounds for median paleointensity.

paleoexposure duration and are most consistent with a history involving steady-state erosion prior to burial. We use SPRITE to calculate the range of plausible steady-state erosion rates to explain the observed ^{21}Ne depth profile, and compare these erosion rates to those originally estimated by Libarkin et al. (2002). To do this, we use a modified version of the exposure age calculation script in SPRITE. We test a range of erosion rates by solving for the concentration of cosmogenic ^{21}Ne over a fixed duration (500 ka) and using time-varying scaling factors. We compare the concentration of cosmogenic ^{21}Ne SPRITE predicts at each sample depth with the measured concentration by calculating a reduced χ^2 misfit statistic.

3. Results

3.1. Effects of Time-Varying Magnetic Field Intensity, Paleolatitude, and Climate-Modulated Paleatmospheric Depth on Scaling Factors

Figure 4 demonstrates that the strength of the geomagnetic field has varied significantly over the last 70 Ma. The present day field strength of $7.7 \times 10^{22} \text{ Am}^2$ (IGRF-11, Finlay et al. (2010)) is 43% higher than the average from 2 to 70 Ma of $4.4 \times 10^{22} \text{ Am}^2$. The highest median field strength in the MCADAM model of $7.5 \times 10^{22} \text{ Am}^2$ occurs at 53.75 Ma, while the lowest median field strength in the MCADAM model of $2.6 \times 10^{22} \text{ Am}^2$ occurs at 47.5 Ma. The average modeled field strength from 2 to 70 Ma is also 19% lower than the average field strength that the LSDn model applies before 2 Ma of $5.4 \times 10^{22} \text{ Am}^2$ (black arrow in Figures 4d and 4e). This difference arises because the long-term average used by LSDn for time periods prior to 2 Ma (hereafter referred to as the LSDn long-term average) is the mean field strength of the GLOPIS-75/PADM2M record (Figure 4e).

To assess how long-term variations in geomagnetic field strength as well as changes in paleolatitude impact production rate scaling factors, Figure 6 shows how scaling factors calculated with SPRITE vary from 2 to 70 Ma for a location in present-day India when changes in geomagnetic field intensity and paleolatitude are and are not accounted for. For an invariant field strength and latitude, we calculate a scaling factor of 0.71; this is equivalent to what LSDn would calculate for exposures before 2 Ma. If only paleolatitude varies and the field intensity is held constant (dashed curve, Figure 6b), scaling factors are lower than this constant field and latitude model 89% of the time. This is because India has been closer to the equator most of the time during the last 70 Ma than it is at present (Figure 2). Conversely, if only the field strength varies, and latitude is held constant (solid curve, Figure 6b), scaling factors are higher than the constant field and latitude model 74% of the time. This is because paleointensities in MCADAM1.b_cosmo are lower than the LSDn long-term average for $\sim 68\%$ of the duration of

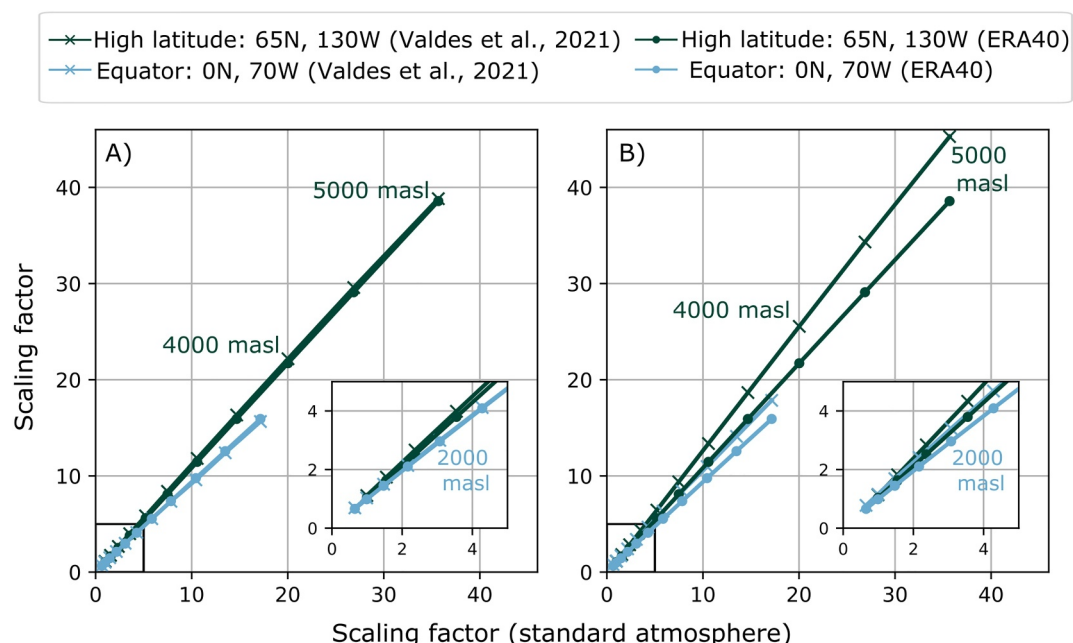


Figure 7. Difference in scaling factor calculated using the standard atmosphere conversion to atmospheric depth versus using the ERA-40-based conversion (circles) or the Valdes-based conversion (crosses) for cosmogenic ^3He production in pyroxene, either (a) at the present day, or (b) at 70 Ma. Note that the scaling factors only change for the Valdes-based conversion between (a) and (b). Blue = 0°N , Green = 65°N . Elevation increments of 500 m are shown with tick marks for elevations ranging from 0 to 5,000 masl.

SPRITE where MCADAM is applied (i.e., 2–70 Ma). When the effects of time-varying field strength and paleolatitude are combined (solid curve, Figure 6a), scaling factors are higher than a constant field and paleolatitude (dashed curve) 63% of the time. This demonstrates that although variations in paleolatitude alone do not change the scaling factor substantially, their effects can become greater when combined with the effects of paleointensity. This is because the dependence of R_c (Equation 4) on paleointensity (M_i) and paleolatitude (λ) is nonlinear. Overall, incorporating both a time-varying paleointensity and paleolatitude make scaling factors higher relative to LSDn due to the field strength being frequently lower than the LSDn long-term average over the last 70 Ma.

In Figure 7, we assess the effect of different atmospheric depth conversion methods on production rate scaling factors. To isolate the effects of the time-varying Valdes-based conversion, we use the LSDn long-term average field intensity and constant latitude for Figure 7a (present day) and Figure 7b (70 Ma). In Figure 7a, scaling factors calculated using the ERA-40-based and Valdes-based conversions are nearly identical; deviations remain small even at high elevations. Both conversions yield slightly lower scaling factors than the standard atmosphere approximation at the equator (blue curves, Figure 7a) and higher scaling factors at a subpolar latitude (green curves, Figure 7a). In comparison, at 70 Ma (Figure 7b) the scaling factors determined using the Valdes-based conversion are consistently higher than those for the present day (Figure 7a) at both subpolar and low latitudes, and the differences between the time-varying Valdes-based conversion and both the ERA-40-based and standard atmosphere conversions become greater with increasing elevation. These results are consistent with the results of Staiger et al. (2007), who found differences in production rate at the LGM relative to today are most pronounced at high altitudes.

In addition to long-term variations in modeled scaling factors, we can evaluate how using SPRITE impacts paleoexposure durations that are estimated from cosmogenic noble gas concentrations. For example, it is possible to measure cosmogenic ^3He concentrations in pyroxene from the flow tops of lava flows in the Deccan Traps in India to determine how long that flow top was exposed before being buried by the next lava flow. The majority of the Deccan Traps erupted between ~ 66.5 and 65.2 Ma (Schoene et al., 2019; Sprain et al., 2019). Within this time period, the paleointensity from MCADAM1.b_cosmo is more than 50% lower than the LSDn long-term average, and the paleolatitude of Deccan Traps lava flows in India during their eruption was $\sim 22^\circ\text{S}$. During the 65–70 Ma

time steps, the Valdes et al. (2021) simulations predict relatively warm temperatures both globally and in the vicinity of the Deccan Traps, which is consistent with proxy data about climate conditions at the end of the Cretaceous (Hull et al., 2020; Li & Keller, 1998; Scotese et al., 2025; Tobin et al., 2017; Wilf et al., 2003). Here, we consider a hypothetical Deccan Traps lava flow, emplaced at sea level and experiencing negligible erosion before burial by the next lava flow that has a cosmogenic ^3He concentration in pyroxene of 2×10^6 atoms/g. For this hypothetical lava flow and cosmogenic ^3He concentration, we calculate a paleoexposure duration with SPRITE of 17.6 Kyr using the standard atmosphere approximation to convert elevation to atmospheric depth. When using the ERA-40-based conversion, we calculate an exposure duration of 18.6 Kyr, which is $\sim 6\%$ older than when using standard atmosphere conversion, while when using the Valdes-based atmospheric conversion, we calculate an exposure duration of 16.5 Kyr, which is 6% younger than when using the standard atmosphere conversion. In contrast, when we use the standard atmosphere conversion but instead of MCADAM use the LSDn long-term average field intensity, we calculate an exposure duration of 22.2 Kyr, which is $\sim 26\%$ longer than we predict with the SPRITE framework.

We can test the permissible range of paleointensities derived from MCADAM, and evaluate its effects on calculated exposure ages for this Deccan Traps example. First we calculate the exposure age using the 25% confidence interval, then the 75% confidence interval (i.e., the prediction bounds presented in Figure 4). Using the 25% confidence interval, the calculated exposure age is $<3\%$ shorter than the exposure age calculated using the median paleointensity, and $<4\%$ longer using the 75% confidence interval. Within the entire interquartile range, the calculated exposure ages are still 21% – 29% younger than using the LSDn long term average. Therefore, even within the prediction bounds of the MCADAM model, the calculated exposure ages are still significantly younger relative to using the LSDn long term average.

3.2. Continuous, Multi-Million Year Exposure Histories of the Pacific Paleosurfaces, Chile

Evenstar et al. (2017) measured cosmogenic ^3He in pyroxene and ^{21}Ne in quartz from 45 clasts from 20 sites across the PPS in Chile. They calculated exposure ages, which they interpret as a minimum estimate for the last time there was fluvial activity occurring at these sites, assuming no subsequent disturbances due to denudation or burial. Based on the ^3He and ^{21}Ne measurements and other lines of geologic evidence, Evenstar et al. (2017) interpreted that there are 5–6 possible paleosurfaces within the PPS with ages of ~ 23 , 10, 11, 7, and 3 Ma that reflect variations in regional sediment supply due to tectonic activity in the presence of long-lasting hyperarid climate conditions. Similarly, Dunai et al. (2005) measured cosmogenic ^{21}Ne in quartz from 16 samples comprising both individual clasts and amalgamated samples from what is inferred to be one of the oldest preserved surfaces in the PPS. The ancient apparent exposure ages reported by Dunai et al. (2005) have been inferred to document the onset of hyperarid conditions in association with Andean tectonic uplift since Oligocene-Miocene times.

Figure 8 highlights the effect of utilizing scaling factors from the SPRITE framework with a time-varying paleointensity and paleolatitude, in the absence of an imposed uplift rate, to calculate apparent exposure ages for these surfaces inferred to have experienced continuous exposure over long (>2 Myr) durations. We compare exposure ages calculated using the SPRITE framework to exposure ages calculated assuming the long-term paleointensity average of the LSDn scaling scheme and a constant latitude. For samples with exposure ages between 2 and 6 Ma, SPRITE predicts exposure ages that are 0.001 – ~ 0.14 Ma, or 0.8% – 2.8% , older than the exposure ages calculated using the LSDn parameterization (Figures 8b and 8c). For samples with exposure ages greater than 6 Ma, SPRITE predicts consistently younger exposure ages than the LSDn parameterization, with the difference in exposure age increasing with increasing exposure age. The differences are most pronounced for exposure ages greater than 10 Ma, where calculated exposure ages can differ by several million years (i.e., $>5\%$; Figures 8b and 8c).

We also calculated exposure ages for these samples, but implemented an uplift rate of 0.004 cm/yr as was assumed in the original studies of these samples (Figure 9) (Dunai et al., 2005; Evenstar et al., 2017). Using both the SPRITE framework and LSDn, the exposure ages calculated with uplift (Figure 9) are consistently older than in the no uplift calculation (Figure 8). This result is expected, because in the uplift scenario the atmospheric depth is greater and therefore the production rate at each site is lower in the geologic past than for the present day, requiring a longer exposure time for the same cosmogenic nuclide abundance to be produced. The trends in exposure ages calculated with uplift imposed are consistent with the trends observed with no uplift: for samples with exposure ages of 6 Ma or younger, we still predict an older exposure age with the SPRITE framework than with LSDn, and for samples with exposure ages older than 6 Ma, we still predict younger exposure ages with

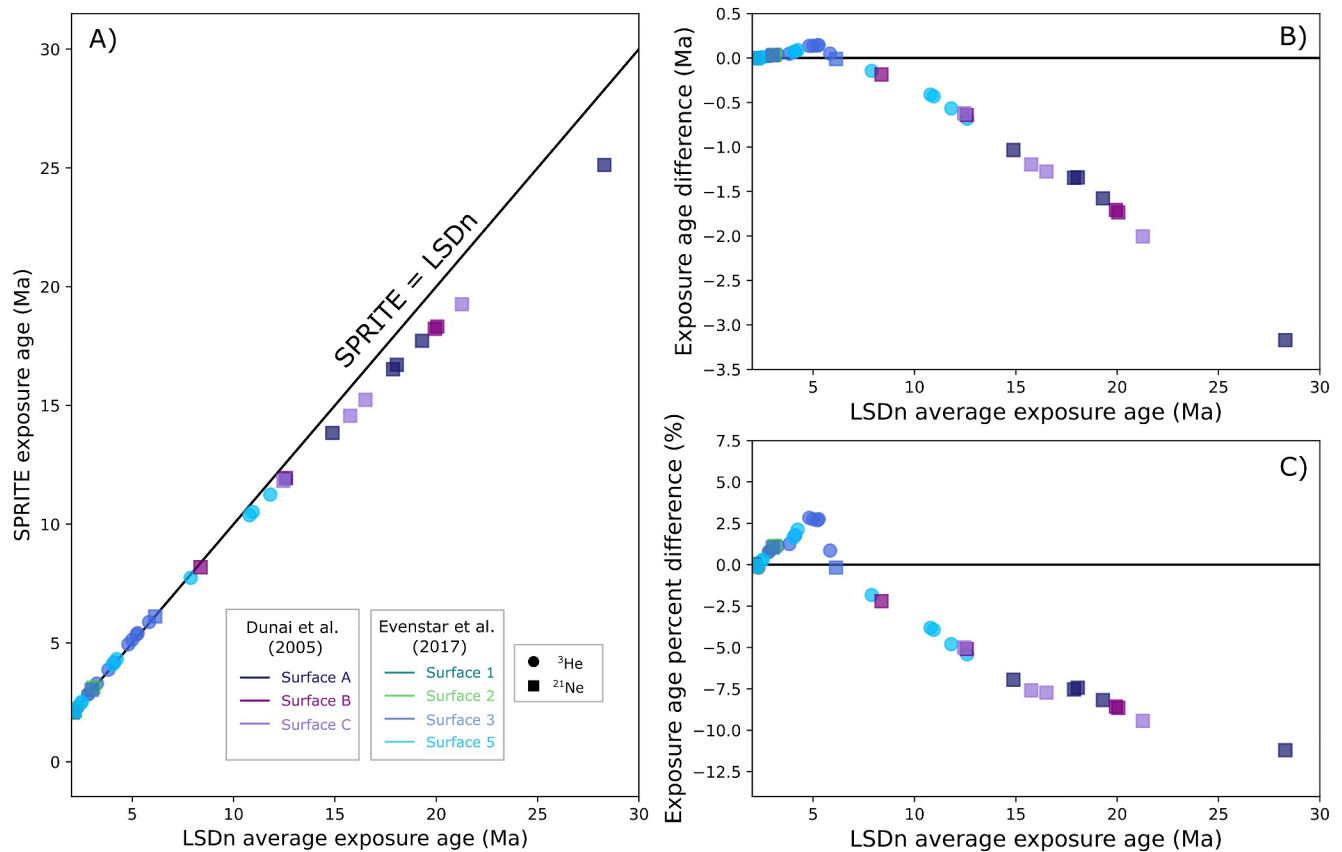


Figure 8. Exposure ages calculated using the cosmogenic noble gas measurements of Evenstar et al. (2017) and Dunai et al. (2005), using a time-varying magnetic field and paleolatitude versus constant latitude and long-term average paleointensity before 2 Ma. For this first calculation we assume that the elevations of the samples have not changed, in order to evaluate the magnitude of the time-varying models within the SPRITE framework on long-duration production rates. (a) Comparison of exposure ages calculated using the long-term field strength average of LSDn to the exposure ages calculated using SPRITE. A 1:1 line is shown for reference. Analytical uncertainties are smaller than symbols plotted. Colors denote distinct surfaces, as defined by Dunai et al. (2005) and Evenstar et al. (2017). Squares show exposure ages based on measurements of cosmogenic ^{21}Ne in quartz, while circles show exposure ages determined from cosmogenic ^3He concentrations in pyroxene. (b) Difference in calculated exposure age between SPRITE and the LSDn long term average field strength. Negative values indicate that SPRITE predicts a younger exposure age than LSDn and vice versa. (c) Percent difference in calculated exposure ages.

SPRITE than with LSDn. However, the difference in exposure ages calculated between the uplift and no uplift scenarios are larger when calculated with LSDn than when calculated with the SPRITE framework. This is because the fixed geomagnetic field strength in LSDn is higher on average than the time-varying field strength in the SPRITE framework, particularly before 10 Ma. The lower paleo-field strength applied in the SPRITE framework essentially counteracts the effects of the greater atmospheric depths imposed in the uplift scenario on production rates, resulting in a smaller difference in cumulative exposure time required to produce the observed cosmogenic nuclide abundance. As a consequence, when we compare the exposure ages calculated with the SPRITE framework versus LSDn in the uplift scenario, both the absolute and percent differences in exposure ages are larger (Figures 9b and 9c) than in the calculations with no uplift (Figures 8b and 8c). Thus, even with a geologically reasonable uplift rate imposed (Dunai et al., 2005), exposure age differences on the order of several million years (i.e., greater than 5%) are still predicted for most samples with exposure ages greater than 10 Ma from these data sets (Figures 9b and 9c).

3.3. Paleoerosion Rate of the Fish Canyon Tuff

Libarkin et al. (2002) measured cosmogenic ^{21}Ne in quartz in the Oligocene Fish Canyon Tuff in Colorado, which was buried by the younger Carpenter Ridge Tuff. Between 15 and 55 cm below the Fish Canyon Tuff-Carpenter Ridge Tuff contact, Libarkin et al. (2002) measured an ancient cosmogenic ^{21}Ne component that accumulated in the surface of the Fish Canyon Tuff prior to its burial by the Carpenter Ridge Tuff. Having an independent

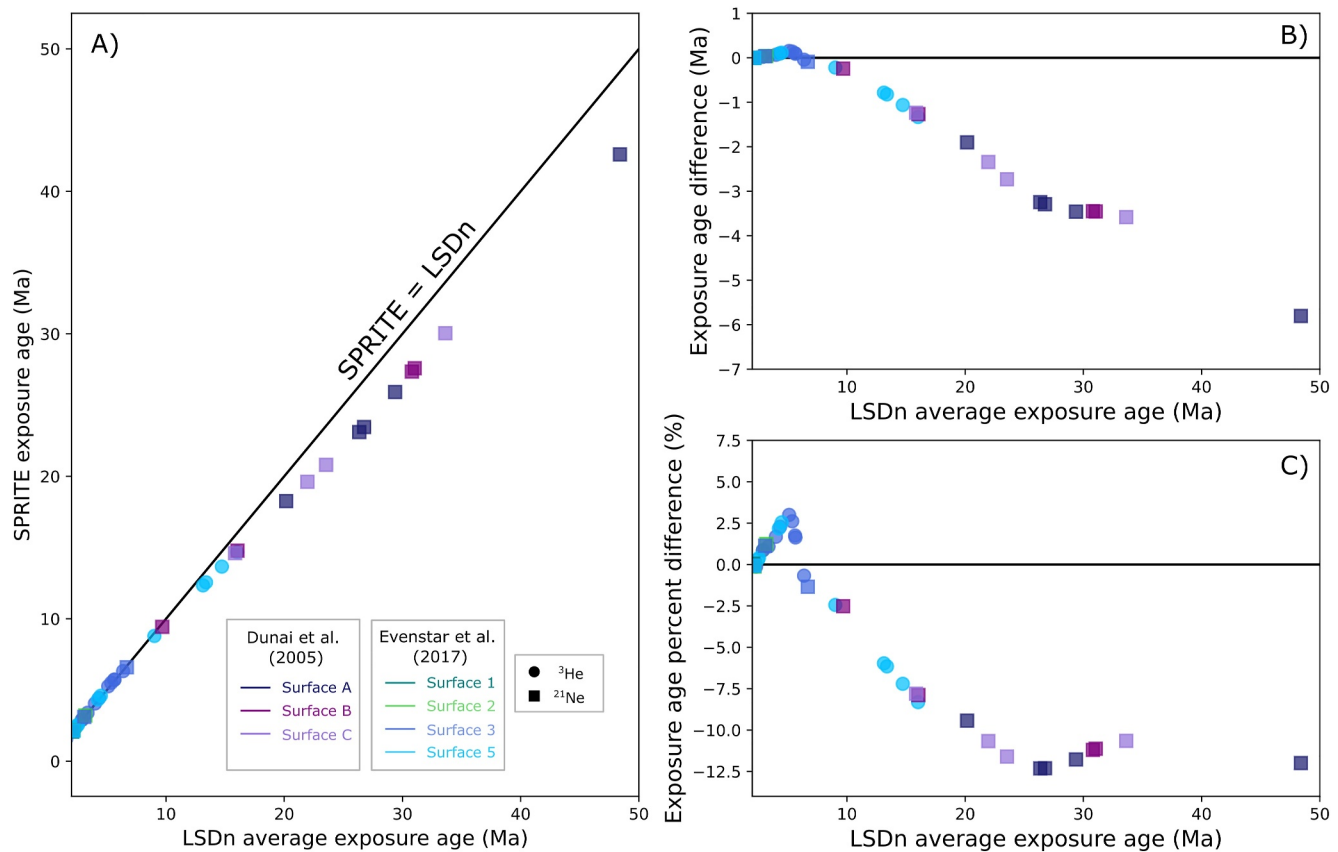


Figure 9. Exposure ages from Evenstar et al. (2017) and Dunai et al. (2005), re-calculated using a time-varying magnetic field (before 2 Ma) and paleolatitude versus an average paleointensity (before 2 Ma) and constant latitude. Exposure ages for both sets of samples are calculated using an uplift rate of 0.004 cm/yr after Dunai et al. (2005). (a) Comparison of exposure ages calculated using the long-term field strength average of LSDn to the exposure ages calculated using SPRITE. A 1:1 line is shown for reference. Analytical uncertainties are smaller than symbols plotted. Colors and symbols are as plotted in Figure 8. (b) Difference in calculated exposure age between SPRITE and the LSDn long term average field strength. Negative values indicate that SPRITE predicts a younger exposure age than LSDn and vice versa. (c) Percent difference in calculated exposure ages.

estimate of the hiatus duration between the Fish Canyon and Carpenter Ridge Tuff eruptions from $^{40}\text{Ar}/^{39}\text{Ar}$ geochronology of ~500 ka, Libarkin et al. (2002) use their observed ancient cosmogenic ^{21}Ne concentrations to quantify paleoerosion rates of the Fish Canyon Tuff following its emplacement.

We use SPRITE to calculate the concentration of cosmogenic ^{21}Ne as a function of depth following the eruption of the Fish Canyon Tuff at 28.201 Ma (Kuiper et al., 2008) assuming an exposure duration of 500 ka and either the present-day elevation of 2,500 masl during exposure (Figure 10a) or a paleoelevation at sea level (Figure 10b). Note that the erosion rates we predict are not sensitive to the exact duration or start time (i.e., eruption age for the Fish Canyon Tuff), as the modeled cosmogenic ^{21}Ne concentration reaches steady state within 40,000 years from the start of each magnetic field model time bin. We find that a paleoerosion rate of 5.5×10^{-3} cm/yr or 3×10^{-4} cm/yr (at 2,500 masl or sea level, respectively) minimizes the reduced χ^2 misfit statistic between the measured and predicted concentrations of cosmogenic ^{21}Ne in this depth profile. This erosion rate is very similar to the erosion rate determined using the long-term magnetic field strength of LSDn of 5.4×10^{-3} cm/yr or 2.9×10^{-4} cm/yr (at 2,500 masl or sea level, respectively). Our calculated paleoerosion rate suggests that ~30–150 m of the Fish Canyon Tuff were eroded away, at least locally, prior to the eruption of the Carpenter Ridge Tuff.

4. Discussion

4.1. Temporal Variations in Geomagnetic Field Strength

The paleomagnetic record shows that time-averaged magnetic field strength was lower for much of the Cenozoic compared to the past 2 Myr (e.g., Biggin et al., 2009, 2015; Bono et al., 2021, 2022; Kulakov et al., 2019; Selkin &

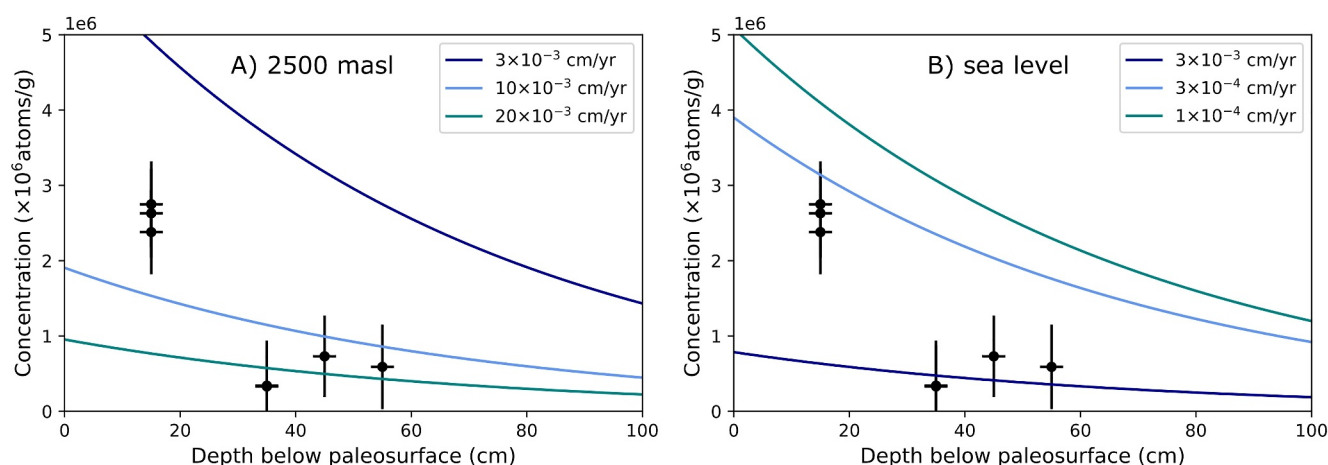


Figure 10. Paleoerosion rates calculated with SPRITE for the Oligocene Fish Canyon Tuff before it was buried by the Carpenter Ridge Tuff ~ 500 ka later, compared to measured concentrations of cosmogenic ^{21}Ne in quartz reported by Libarkin et al. (2002). Calculations assume (a) the present-day elevation of the Fish Canyon Tuff where it was sampled (2,500 m), or (b) emplacement at sea level, after Libarkin et al. (2002). A paleoerosion rate of $5.5 \times 10^{-3} \text{ cm/yr}$ at 2,500 masl or $3 \times 10^{-4} \text{ cm/yr}$ at sea level minimizes the reduced chi-square misfit between the observed and predicted cosmogenic ^{21}Ne concentrations.

Tauxe, 2000). As such, the MCADAM1.b_cosmo model incorporated in SPRITE yields lower paleointensities, on average, from 2 to 70 Ma compared to other scaling models like LSDn, which for time periods >2 Ma uses a constant value derived from the long term average value of recent data (Figure 4). These lower paleointensities permit a higher flux of incoming cosmic-ray particles at all subpolar latitudes, which results in larger production rate scaling factors and ultimately younger exposure ages on average calculated by SPRITE versus other scaling schemes, given the same measured cosmogenic nuclide abundance.

Our results demonstrate that the temporal variations in field strength incorporated into MCADAM1.b_cosmo have a nontrivial effect on the cosmogenic nuclide production rates that SPRITE calculates, particularly for relatively short duration paleoexposures of geologic surfaces during intervals of time when the field strength was significantly different from the LSDn long-term average. In both our hypothetical calculation for a buried Deccan Traps lava flow and our calculations using existing cosmogenic noble gas data sets (Figures 8 and 9), we find that the effects of variations in the geomagnetic field intensity on cosmogenic production rate are often significantly larger than the effects of changes in paleolatitude or climate-driven changes in paleoatmospheric depth.

It is worth noting that the paleointensity record has decreasing variance back in time due to data sparsity that captures less of the magnetic field's shorter term changes. This sparsity is due in large part to the fact that there is a lack of well-preserved outcrops with robust paleointensity records identified for many periods of geologic time, leading to gaps in the record. This sparsity is also partially due to the fact that paleointensity is a difficult component of the magnetic field to constrain due to time-intensive measurements, experimental alteration, and low success rates. The addition of new paleointensity data in the future will likely improve the MCADAM1.b_cosmo model's ability to resolve shorter duration (i.e., sub-Ma) variations in the geomagnetic field strength. However, these short-duration variations in the field strength are typically smaller in magnitude than the longer-term evolution of the field strength over millions of years. For example, this can be demonstrated by comparing a 1 Myr interval in Figure 4e to the magnitude of variations captured from 0 to 70 Ma in Figure 4d. The observation that the Earth's magnetic field strength has been, on average, lower over the last 70 Myr than the present-day field is a robust feature of the MCADAM model, as are the secular trends in paleointensity observed on timescales of Myr and longer. This implies that production rates of cosmogenic nuclides have been higher over many intervals of the Cenozoic than we would infer from late Quaternary cosmogenic nuclide data sets (Ackert et al., 2003; Cerling & Craig, 1994; Licciardi et al., 2006, 2008; Poreda & Cerling, 1992) and production rate scaling models built for the late Quaternary (e.g., Lifton et al. (2014)).

4.2. Temporal Variations in Atmospheric Depth

The effects of climate-driven temporal variations in atmospheric depth on cosmogenic production rates have not been implemented in existing scaling models. We incorporated an option for accounting for such climate-driven

variations in SPRITE using the model simulations of Valdes et al. (2021). We emphasize that these simulations have a coarse temporal resolution (i.e., 5 Ma time bins compared to the 250 ka resolution of MCADAM1.b_cosmo as implemented here). While sometimes the Valdes-based conversions of elevation to atmospheric depth implemented in SPRITE yields greater atmospheric depths, we have found that most often the Valdes-based conversion yields smaller atmospheric depths than those calculated with the standard atmosphere approximation or the ERA-40 based conversion for the same location (Figure 7), which results in higher production rates when using the Valdes-based conversion relative to the other two methods. As more climate model simulations spanning the Cenozoic and earlier become available, we will be able to evaluate if the lower atmospheric depths we often obtain using the Valdes-based conversion are a robust feature, similar to how we can infer that the lower average paleointensity we obtain over the last 70 Ma from the MCADAM model is a robust feature. We found that variations between the three approaches for determining atmospheric depth typically yield <10% differences between calculated scaling factors, while differences that can result from changes in paleointensity or combined changes in paleointensity and paleolatitude can greatly exceed 10%. For example, in our hypothetical Deccan Traps scenario, the difference in exposure age is ~6% between all three approaches for determining atmospheric depth. However, the difference between exposure age is greater than 25% when only considering paleointensity and paleolatitude variations. Similarly, we observe greater than 10% differences in exposure age for paleosurfaces in the Atacama Desert for samples older than ~20 Ma (Figures 8 and 9) when only considering variations in paleointensity and paleolatitude. Given the coarseness of the Valdes-based conversion and the relatively smaller relative effects of the specific atmospheric conversion on production rates compared to the potential paleointensity effects, we therefore focused on the combined effects of paleointensity and paleolatitude on production rates in our evaluations of existing cosmogenic noble gas data sets.

Changes in a sample's paleoelevation and therefore atmospheric depth due to tectonic processes are arguably more important and harder to constrain than climate-driven changes in atmospheric depth, and this remains an outstanding challenge in applications of cosmogenic nuclides to deeper timescale studies. At present, the SPRITE framework allows the user to provide an uplift or subsidence rate (cm/yr) if they have independent constraints on what such rates might be, in the case of a continuous exposure history, or to impose a paleoelevation, in the case of a paleoexposure episode. Should local or regional paleoelevation models become available, these could be incorporated into instances of SPRITE for a more robust treatment of paleoelevation effects. Alternatively, the SPRITE framework may be used to predict time-dependent scaling factors for different hypothesized paleoelevation histories and, given if other constraints on a sample's surface exposure history can be supplied, used for paleoaltimetry (e.g., Blard et al., 2006, 2019). Finally, we note that, in the absence of robust paleoelevation constraints for paleoexposures, the SPRITE framework can still be utilized to obtain relative exposure age or erosion rate differences between geologic surfaces with a known vertical separation, such as different lava flows within the Deccan Traps, because their paleoelevation and present-day elevation differences will be the same. The relative difference in exposure ages, when paired with absolute chronological constraints on the emplacement of lava flows, could enable constraints on the paleoexposure durations (or paleoerosion rates) and paleoelevations to be jointly obtained.

4.3. Impact of SPRITE on Estimates of Paleoexposure Duration and Paleoerosion Rates

Paleoexposure histories recorded on geologic surfaces are typically short relative to the changes in paleolatitude and magnetic field intensity that we incorporate into SPRITE. As a consequence, the impact of using SPRITE to determine paleoexposure durations and paleoerosion rates, as opposed to using an existing scaling model such as LSDn with a fixed field intensity and present-day latitude, can vary depending on when and where a geologic surface's paleoexposure history took place. For example, we demonstrated that for the paleoexposure of a Deccan Traps lava flow with a hypothetical cosmogenic ^3He concentration of 2×10^6 atoms/g in pyroxene, we would predict a paleoexposure duration with LSDn that is ~26% longer than what we calculated with SPRITE. This paleoexposure duration difference is primarily driven by the fact that the paleointensity in MCADAM1.b_cosmo during the eruption of the Deccan Traps at ~66 Ma is more than 50% lower than the LSDn long-term average field strength. Paleolatitude effects on exposure age in this case are secondary to the paleointensity effects but not negligible. Using the long term average paleointensity from LSDn and varying the site paleolatitude, the LSDn paleoexposure duration is ~13% longer than what we calculate with SPRITE. This is because around 66 Ma, the paleolatitude of the site was ~25°S, whereas the present day latitude is 20°N. The relative effects of paleointensity and paleolatitude are highlighted in Figure 6b. When only paleolatitude changes (dashed curve), the scaling factor

at 66 Ma is 1.08 times larger than the present-day scaling factor; however, when only paleointensity changes (solid curve), the scaling factor at 66 Ma is ~ 1.3 times larger than the present-day scaling factor.

We can compare this example from the Deccan Traps, where the use of SPRITE does have a significant impact on the calculated paleoexposure duration, to an analogous example of paleoexposure of a lava flow in the Columbia River Flood Basalts. Over 95% of volcanism from the Columbia River Flood Basalts (CRFB) erupted between 16.7 and 15.9 Ma (Kasbohm & Schoene, 2018). During this time interval, the MCADAM1.b_cosmo paleointensities are comparable to or slightly higher than the LSDn long-term average. If we consider the same hypothetical sample as we did for the Deccan Traps with 2×10^6 atoms/g of ^3He in pyroxene with negligible erosion and at sea level, with SPRITE we calculate an exposure duration of 17 Kyr. Using the same input parameters but using the LSDn long-term average intensity and present day latitude, we calculate an exposure duration of 16.8 Kyr. Between today and 16.7 Ma, the paleolatitude of the CRFB has not changed significantly, so the difference in calculated exposure age is almost entirely due to the small difference in paleointensity between MCADAM1.b_cosmo for this time period and the LSDn long term average. However, the difference in these ages is an order of magnitude less than the difference we predicted for the same scenario in the Deccan Traps, and would be well within the analytical uncertainty of a cosmogenic ^3He measurement. We see essentially the same effect for the paleoerosion rates that we estimate for the cosmogenic ^{21}Ne data set from the Fish Canyon Tuff reported by Libarkin et al. (2002). The paleolatitude and present-day latitude of the Fish Canyon Tuff are very similar, and the average magnetic field strength between ~ 28.2 and ~ 27.5 Ma from MCADAM1.b_cosmo is $\sim 5.65 \times 10^{22} \text{ Am}^2$, which is very similar to the long-term average assumed by LSDn of $5.5 \times 10^{22} \text{ Am}^2$. As a result, the paleoerosion rates we estimate using the cosmogenic ^{21}Ne measurements of Libarkin et al. (2002) are indistinguishable using SPRITE and using LSDn.

In summary, the effects of paleolatitude on production rates and therefore on paleoexposure durations or paleoerosion rates are secondary, and sometimes negligible, relative to the effects of paleointensity in SPRITE (e.g., Figure 7b). While there are several periods of time during the last 70 Ma (such as at the ends of the early Miocene and early Oligocene) when the magnetic field intensity was similar to the long-term average field intensity used by LSDn for exposures before 2 Ma, there are also many time periods during which the field strength was significantly different, such as during the eruption of the Deccan Traps at ~ 66 Ma (Figure 4). During these intervals of time, accounting for the difference in paleointensity can have a nontrivial impact on cosmogenic nuclide production rates (e.g., Figure 6) and therefore the paleoexposure durations or paleoerosion rates determined from cosmogenic nuclide measurements.

4.4. Impact of SPRITE on Continuous, Multi-Million Year Exposure Histories

Production rates for continuous, multi-million year exposure histories, like those experienced by the samples studied by Dunai et al. (2005) and Evenstar et al. (2017), integrate changes in paleointensity and paleolatitude through time. This contrasts with the examples above of relatively short paleoexposure histories for which we essentially need snapshots of the paleointensity, paleolatitude, and paleoatmospheric depth at a particular point in the geologic past to determine production rates. For continuous exposure histories, this time integration means that the long-term average of paleointensity and paleolatitude will impact calculated production rates and therefore exposure ages. For example, the difference between apparent exposure ages we calculate using SPRITE and the LSDn long-term average are within the reported analytical uncertainties for exposure ages younger than 10 Ma from the Dunai et al. (2005) and Evenstar et al. (2017) data sets, whether or not an uplift rate is utilized (Figures 8a and 9a). As the apparent exposure ages increase for samples from the Pacific Paleosurfaces to greater than 10 Ma, the difference in exposure ages calculated using the two models becomes significantly greater than analytical uncertainties, with maximum exposure age differences exceeding 10% for those samples with the oldest apparent exposure ages (Figures 8c and 9c). This increased difference in exposure age predominantly results from the paleointensity from MCADAM1.b_cosmo being lower than the LSDn long term average $\sim 80\%$ of the time between 10 and 35 Ma (Figure 4). In the model of Besse and Courtillot (2002), over the last 70 Ma, South America has only moved $\sim 2\text{--}3^\circ$ south relative to its modern day position (Figure 2b). The difference in exposure ages we calculate are therefore due almost entirely to variations in paleointensity, rather than paleolatitude, for this example. Paleolatitude effects may be more significant for surfaces experiencing continuous, multi-million year exposure histories on other continents, such as Australia (Fujioka & Chappell, 2011).

Dunai et al. (2005) and Evenstar et al. (2017) include an elevation increase of 40 m/Ma into their production rate calculations in order to calculate apparent exposure ages. This uplift rate results in an increase in production rates toward the present day due to a decrease in the atmospheric depth, and therefore older apparent exposure ages (Figure 9) than we calculate in the no uplift scenario (Figure 8), both utilizing the SPRITE framework and the original LSDn scaling model. The difference between exposure ages calculated using the SPRITE framework relative to LSDn scaling in the uplift scenario are greater than the no uplift scenario (Figure 9c), particularly for samples >10 Ma. This is because, when the samples resided at lower elevations further back in time, applying the long term average paleointensity used in LSDn results in even lower production rates, and therefore requires increasingly longer exposure durations. When we use SPRITE, the paleointensity also decreases further back in time, essentially counterbalancing the effects of the lower elevations. Overall, the effects of implementing the SPRITE framework for continuous exposures are most pronounced with paleosurfaces greater than ~20 Ma. In summary, we found that when integrating lower paleointensities over >10 Ma timescales, the differences in exposure ages calculated with the SPRITE framework relative to LSDn become significant (Figures 8 and 9). These effects become even more pronounced when external factors (e.g., uplift rate) enhance the differences between using a time-varying versus constant long-term average paleointensity (Figure 9).

4.5. Use Cases and Future Improvements

We designed the SPRITE framework to estimate production rates prior to the Quaternary and up to 70 Ma. SPRITE is complementary to and builds upon the previous body of work determining production rates and scaling factors, but takes the step of incorporating deeper time information about geomagnetic field intensity, paleolatitude, and paleoclimate to improve our ability to interpret cosmogenic nuclide data beyond the Quaternary. SPRITE will be particularly applicable in the following geologic situations:

- Short exposure durations (<<250 Kyr) that take place before the Quaternary. Individual 250 ka intervals in SPRITE can have a paleointensity value that is more than 50% different from a long-term average value or than the average paleointensity over the Quaternary. Examples of applications include constraining hiatus durations between ancient lava flows, determining paleoerosion rates (e.g., Libarkin et al., 2002), and quantifying sediment transport and storage timescales in the geologic past (e.g., Sinclair et al., 2019).
- Long (many Myr) continuous exposure durations at any point in the last 70 Ma. Over longer exposure durations, time-integrated paleointensities have a higher probability of being lower, and production rates are more likely to be higher than the average paleointensity production rate over the Quaternary. An example of this application is quantifying the exposure duration of a paleosurface (e.g., the Pacific Paleosurface in the Atacama Desert) to determine how long it has been stable with respect to its regional climatic and tectonic history (Dunai et al., 2005; Evenstar et al., 2009, 2017; Ritter, Binnie, et al., 2018; Ritter, Stuart, et al., 2018).
- Paleoexposure and/or paleoerosion rate estimates from sites in India, Eurasia and Australia. These regions have experienced significant (>10°) variations in latitude over the last 70 Ma, meaning that production rates calculated using present-day latitudes may be significantly different from production rates calculated using paleolatitudes, depending on when the period of exposure took place.

SPRITE can be improved upon in the future as new data and models become available, and the framework is designed to be straightforward to update. For example, the current implementation of muogenic production of cosmogenic ³He is based on the empirical fitting of a ³He depth profile in the CRFB (Larsen et al., 2021), for which significantly different surface contributions and attenuation lengths were found compared to other cosmogenic nuclides (e.g., Balco, 2017). Future work to develop a theoretical model that tests the empirical results of Larsen et al. (2021) could be incorporated into the muogenic ³He production rate calculations in the SPRITE framework. A theoretically grounded muogenic production model for ³He could enable new applications involving deep (~tens of meters) depth profiles in presently exposed surfaces to estimate long-term exposure durations and erosion rates. Another anticipated future improvement would be the incorporation of paleoclimate model simulations at a higher temporal resolution than the simulations of Valdes et al. (2021) in order to model climate-driven changes in atmospheric depth, as well as regional paleoelevation models to account for changes in atmospheric depth driven by tectonic processes, should such models become available. Finally, additions to the global paleointensity record can be readily incorporated into MCADAM1.b_cosmo to improve the paleomagnetic effects on production rates as such data are generated.

5. Conclusions

Existing cosmogenic nuclide production rate scaling schemes were optimized for the late Quaternary. We developed the SPRITE framework to account for processes that influence cosmogenic nuclide production rates on longer geologic timescales. The SPRITE framework extends the LSDn scaling model to 70 Ma by incorporating (a) site-specific changes in paleolatitude, (b) a geomagnetic field intensity model rooted in paleointensity data available over Earth's history, and (c) climate-driven atmospheric depth variations from a suite of GCM simulations. We demonstrate with example data sets from the literature as well as model calculations that incorporating these effects into SPRITE can lead to nontrivial changes in cosmogenic nuclide production rates, which in turn can lead to differences in calculated exposure durations or erosion rates. The effects of paleolatitude and climate-driven changes in paleoatmospheric depth are generally secondary relative to the effects of paleointensity. Paleolatitude changes can, however, have strong local effects at sites which experienced significant plate motions within the last 70 Ma, and climate-driven changes on paleoatmospheric depth become more important at high elevations (>3 km). Over many time periods in the last 70 Ma, the magnetic field strength was lower than the present day as well as lower than the long-term average field strength that LSDn assumes before 2 Ma. As a consequence, cosmogenic nuclide production rates were higher during these intervals of the geologic past. Accounting for these secular changes in paleointensity can result in substantially different estimates of exposure duration, particularly for paleoexposure histories. As the number of “deep time” applications of stable cosmogenic nuclides continues to grow, the SPRITE framework can be utilized to quantify production rate scaling factors using the approaches developed for late Quaternary studies while accounting for factors that are uniquely important on these longer timescales.

Data Availability Statement

Version 4 of SPRITE used for scaling factor and exposure age calculations is published via Mijum and Bristol (2024).

Acknowledgments

This work was supported by NSF Awards 2016748 to MMT and 2016763 to CJS. MMT was also supported by an Alfred P. Sloan Research Fellowship, and CJS was also supported by NSF Award 2054605, 2245629, 2321344, and 2237807. We thank G Balco for helpful discussions and access to existing cosmogenic production rate code libraries. We also thank two anonymous reviewers and AE P. van der Beek for helpful feedback that significantly strengthened this work.

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