

## SUPPLEMENTARY METHODS

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### 1. Sample preparation and quartz isolation.

We collected soil samples using an auger having a bucket with inside diameter of 7.8 cm. For each sample, the auger was advanced 7-12 cm and the beginning and ending vertical positions were directly measured on the auger shaft. Each sample was then weighed to calculate soil density. Bulk densities ranged from 0.93 g/cm<sup>3</sup> in the A horizon to 2.2 g/cm<sup>3</sup> at depth. The average density in the 170-cm mixed layer that we use later in the model calculations is 1.7 g/cm<sup>3</sup>. We then separated quartz from the samples by disaggregating and wet-sieving to a 0.25-0.5 mm grain size range, followed by cleaning in water in a sonic bath, soaking in warm dilute HCl to remove iron oxides, and repeated rinsing. This resulted in a sample composed primarily of residual quartz grains remaining after advanced weathering of the soil. We etched the remaining material in 5% HF at room temperature until the quartz appeared visibly clean.

Although external heat was not applied during sonic cleaning, both sonic cleaning and soaking in warm HCl heated samples above room temperature, which could cause loss of helium from quartz. Although of course this could easily be prevented by performing all sample preparation at room temperature, our original intention was to measure only <sup>21</sup>Ne (which is not lost at the temperatures involved). <sup>3</sup>He measurements were only carried out after we obtained the <sup>21</sup>Ne results. Although it is not possible to calculate <sup>3</sup>He loss exactly without knowledge of the entire thermal history of the sample prior to collection, we performed this calculation with limiting assumptions (complete retention of all <sup>3</sup>He produced during the exposure history, followed by 24 hr at 60° C, with <sup>3</sup>He diffusivity in quartz as measured by Shuster and Farley, 2005) and found that no more than 6% of <sup>3</sup>He present would be expected to be lost during sonic treatment. As the limiting assumption of complete retention of all <sup>3</sup>He produced during exposure is unrealistic, actual loss is expected to be much less than 6%. As this is small compared to the uncertainty in the <sup>3</sup>He concentration measurements, we did not make any correction for <sup>3</sup>He loss during sample preparation.

### 2. <sup>3</sup>He and <sup>21</sup>Ne measurements.

We measured concentrations of He and Ne isotopes in quartz samples by noble gas mass spectrometry (NGMS) on the "Ohio" NGMS system at the Berkeley Geochronology Center. 140-170 mg aliquots of clean quartz were encapsulated in Ta packets and heated under vacuum by an 810 nm diode laser. Released gases were then purified by reaction with SAES getters at room temperature, followed by cryogenic separation of remaining noble gases on an activated charcoal trap and introduction of the desired gas into the mass spectrometer. The mass spectrometer is a MAP-215-50 sector field mass spectrometer with updated detectors and electronics. <sup>3</sup>He and all Ne isotopes were measured using a Channeltron-type multiplier operated in pulse-counting mode.

He and Ne were measured on separate aliquots. He was extracted in one heating step of 15 min at 800° C, followed by freezing of He onto the cryotrap at 12.5° K and release at 33° K. As discussed by, e.g., Balco et al. (2016) and Gribenski et al. (2022), this heating schedule has been found to completely degas <sup>3</sup>He from quartz, but does not completely extract <sup>4</sup>He. The most likely reason for this is that some <sup>4</sup>He is contained in more retentive, U/Th-rich, inclusions. This study is only concerned with cosmogenic <sup>3</sup>He, so

we did not continue heating to higher temperature to quantitatively measure total  $^4\text{He}$ , and  $^4\text{He}$  concentrations are not reported. Further details of  $^3\text{He}$  measurements at BGC have been recently reported in Gribenski et al. (2022) and Tremblay et al. (2019). Full process blanks measured by heating empty Ta packets interspersed with the Cougar Pass samples were  $25000 \pm 28000$  atoms  $^3\text{He}$ . Uncertainties on  $^3\text{He}$  concentrations include measurement uncertainty derived from counting statistics, an uncertainty on machine sensitivity derived from replicability of measurements on helium gas standards, and uncertainty in blank subtraction, added in quadrature.  $^3\text{He}$  concentrations are reported in Table S1.

Ne was extracted in two heating steps of 15 min at  $850^\circ\text{C}$  and 12 min at  $1100^\circ\text{C}$ , with separation from He and Ar by freezing of Ne onto the cryotrap at  $33^\circ\text{K}$  and release at  $70^\circ\text{K}$ . Mass-spectrometric measurement of Ne isotopes involves use of a  $^{39}\text{Ar}$  spike to correct for  $^{40}\text{Ar}^{++}$  observed on mass 20, as described by Balco and Shuster (2009). After correction of  $^{20}\text{Ne}$ ,  $^{21}\text{Ne}$ , and  $^{22}\text{Ne}$  signals for interferences, mass discrimination, and mass spectrometer background, we computed excess  $^{21}\text{Ne}$  relative to atmosphere as  $N_{21,\text{xs}} = N_{21,\text{total}} - (N_{20,\text{total}} R_{21/20,\text{a}})$ , where  $N_{21,\text{xs}}$  is excess  $^{21}\text{Ne}$ ,  $N_{20,\text{total}}$  and  $N_{21,\text{total}}$  are the total measured amounts of  $^{20}\text{Ne}$  and  $^{21}\text{Ne}$ , and  $R_{21/20,\text{a}}$  is the atmospheric  $^{21}\text{Ne}/^{20}\text{Ne}$  ratio (0.002959). As we have found that small quantities of Ne released from heating empty Ta packets have atmospheric composition, correction for any  $^{21}\text{Ne}$  blank derived from sample heating is included in the calculation of excess  $^{21}\text{Ne}$  and not computed separately. Uncertainties in excess  $^{21}\text{Ne}$  concentrations are derived from counting uncertainties on mass 21, uncertainty on the  $^{20}\text{Ne}$  signal derived from correction of the mass 20 signal for the  $^{40}\text{Ar}^{++}$  contribution, and an uncertainty in machine sensitivity derived from replicability of measurements on atmospheric Ne standards, added in quadrature. Table S1 reports summary excess  $^{21}\text{Ne}$  concentrations, and Table S2 shows the full step-degassing results for all Ne isotopes. For this work we have not attempted to separate observed excess  $^{21}\text{Ne}$  into separate cosmogenic and nucleogenic inventories, and instead simply refer to ‘inherited  $^{21}\text{Ne}$ ,’ which could be either cosmogenic or nucleogenic.

### 3. $^{10}\text{Be}$ measurements.

Quartz remaining after  $^3\text{He}$  and  $^{21}\text{Ne}$  measurements was further cleaned by additional HF-etching in a sonic bath, and digested in concentrated HF with 0.2 mg of  $^9\text{Be}$  added from carrier obtained from the University of Washington (UW-TRCH-B-1-5;  $^{10}\text{Be}/^9\text{Be} = 0.6 \pm 0.1 \times 10^{-15}$ ). After removal of Si by evaporation of  $\text{SiF}_6$ , beryllium was extracted via column chromatography and pH-controlled hydroxide precipitations following standard protocols (Stone et al., 2004; Ditchburn and Whitehead, 1994). Hydroxides were then converted to  $\text{BeO}$  by heating, mixed with niobium powder, and loaded into targets for  $^{10}\text{Be}/^9\text{Be}$  measurement by accelerator mass spectrometry at the Center for Accelerator Mass Spectrometry, Lawrence Livermore National Laboratory. Ratio measurements were normalized to the 07KNSTD3110 standard having  $^{10}\text{Be}/^9\text{Be} = 2850 \times 10^{-15}$  (Nishiizumi et al., 2007). A full process blank measured with the samples in this study contained  $10900 \pm 1900$  atoms  $^{10}\text{Be}$ ; for all samples, this blank represented  $<2\%$  of total atoms counted. Reported uncertainties in  $^{10}\text{Be}$  concentration include AMS measurement uncertainty derived from counting statistics and target reproducibility, uncertainty in the Be concentration of the carrier, and uncertainties in the blank correction, added in quadrature.  $^{10}\text{Be}$  concentrations are shown in Table S1.

### 4. Random walk model for $^{10}\text{Be}$ and $^{21}\text{Ne}$ production and loss in a mixed soil profile

The forward model used to generate Figure 3 in the main text considers a large number of individual quartz grains that move vertically by a random walk within the soil mixed layer. We considered 5000 individual grains and assumed that the mixed layer extends from the soil surface to 170 cm depth and there is no loss of soil particles at the surface (which is equivalent to specifying zero erosion) or through the bottom boundary of the mixed layer.

Particle depths at the beginning of the simulation are drawn from a uniform random distribution between 0 and 170 cm. Given particles  $i = 1 \dots n$  and timesteps  $j = 0 \dots m$  with timestep duration  $\delta t$ ,  $Z_{ij}$  is the depth of particle  $i$  below the surface in timestep  $j$ , and  $Z_{ij} = Z_{i,j-1} + k$ , where  $k$  is drawn from a random normal distribution with mean 0 and standard deviation  $\sigma_k$  (cm). The impermeable boundary conditions at the top and bottom of the mixed layer are implemented such that grains “bounce” off the boundaries: if  $Z_{i,j-1} + k < 0$ , then  $Z_{ij} = |Z_{i,j-1} + k|$  and if  $Z_{i,j-1} + k > 170$ , then  $Z_{ij} = 170 - (Z_{i,j-1} + k - 170)$ .

For simplicity, we assume that cosmogenic-nuclide production is exponential such that the  $^{10}\text{Be}$  production rate is  $P_{10}(Z) = P_{10}(0)\exp(-Zp/\Lambda)$ , where  $P_{10}(Z)$  is the production rate (atoms/g/yr) at depth  $Z$  (cm),  $P_{10}(0)$  is the surface production rate (atoms/g/yr, estimated using the scaling model of Stone, 2000, as implemented in version 3 of the online exposure age calculator described by Balco et al., 2008 and subsequently updated),  $p$  is the average soil density in the mixed layer ( $1.7 \text{ g/cm}^3$ , derived from field measurements as described above), and  $\Lambda$  is an effective attenuation length for spallogenic production ( $150 \text{ g/cm}^2$ ). The site has negligible topographic shielding and does not experience significant snow cover, so the production rate is not further corrected. The  $^{10}\text{Be}$  concentration in particle  $i$  at the end of the model run is:

$$\sum_j P_{10}(Z_{i,j})\delta t \exp(-\lambda_{10}\delta t(m-j))$$

where  $\lambda_{10}$  is the  $^{10}\text{Be}$  decay constant ( $4.99 \times 10^{-7} \text{ yr}^{-1}$ ). We then compute the average  $^{10}\text{Be}$  concentration of all particles that lie within each 10-cm depth increment at the end of the model run, and these average concentrations are shown by the black line in Figure 3 in the main text. Given the simplifying assumptions of zero erosion and an impermeable mixed layer boundary,  $^{10}\text{Be}$  concentrations below the mixed layer are simply  $(P_{10}(Z)/\Lambda)(1-\exp(-\lambda_{10}m\delta t))$ . Simplifying the model by disregarding muon production, as we have done, could result in an inaccuracy if we were to try to obtain an exposure age by fitting model to data, but does not have any effect on the basic mixing/heating dynamics that we aim to explain with a simple model.

The adjustable parameters in the mixing model are the exposure duration ( $m\delta t$ ) and the distribution of the random walk distance per time step ( $\sigma_k\delta t$ ). Given a  $^{10}\text{Be}$  depth profile with a mixed layer, one can adjust the exposure duration to match the mean  $^{10}\text{Be}$  concentration in the mixed layer and then adjust  $\sigma_k\delta t$  to match the variation in  $^{10}\text{Be}$  concentration with depth. A small value of  $\sigma_k\delta t$  will predict a poorly mixed profile that resembles the depth dependence of the production rate, whereas large  $\sigma_k\delta t$  predicts a nearly constant concentration through the mixed layer. For these data, an exposure duration of 425,000 years,  $\sigma_k = 10 \text{ cm}$ , and  $\delta t = 250 \text{ yr}$  closely matches the measured  $^{10}\text{Be}$  concentrations in the mixed layer (Figure 3 in main text).

Production of  $^{21}\text{Ne}$  in each particle in each timestep is computed in the same way as for  $^{10}\text{Be}$ , except that  $P_{21}(Z) = 4.03 P_{10}(Z)$  (Balco et al., 2019). The  $^{21}\text{Ne}$  concentration in particle  $i$  at the end of the model run is:

$$N_{21,inh} + \sum_j P_{21}(Z_{i,j})\delta t L_{i,j}$$

where  $N_{21,inh}$  is an inherited  $^{21}\text{Ne}$  concentration (atoms/g) that could include nucleogenic  $^{21}\text{Ne}$  and/or cosmogenic  $^{21}\text{Ne}$  produced prior to landslide emplacement, and  $L_{i,j}$  is the fractional loss due to wildfire heating of  $^{21}\text{Ne}$  produced in particle  $i$  in timestep  $j$ . To correctly account for heating loss of inherited  $^{21}\text{Ne}$ , it is assumed to have been “produced” in timestep 0.

To compute  $L_{ij}$ , we assume that surface and subsurface wildfire heating is characterized by a surface temperature  $T_0$  and exponential attenuation constant  $\Lambda_T$ , such that  $T(Z) = T_0 \exp(-Z/\Lambda_T)$ . This is an approximation to the results of soil heating models such as that in Richter et al. (2022).  $^{21}\text{Ne}$  diffusivity in quartz is then assumed to follow an Arrhenius law:

$$\frac{D}{a^2} = \frac{D_0}{a^2} \exp\left(\frac{-E_a}{RT}\right)$$

where  $D$  is the diffusion coefficient ( $\text{cm}^2/\text{s}$ ),  $a$  is the grain radius ( $\text{cm}$ ),  $R$  is the universal gas constant ( $8.314 \text{ J}^\circ\text{K}/\text{mol}$ ),  $T$  is temperature ( $^\circ\text{K}$ ), and  $D_0$  ( $\text{cm}^2/\text{s}$ ) and  $E_a$  ( $\text{J}/\text{mol}$ ) are empirically determined, material-specific constants. We use values of  $D_0$  (reported as  $\ln(D_0/a^2) = 16$ ) and  $E_a$  ( $1.54 \times 10^5 \text{ J}/\text{mol}$ ) measured for  $^{21}\text{Ne}$  in quartz by Shuster and Farley (2005). For simplicity, we also assume that quartz grains in the soil are spheres with the same diameter as the grain analyzed by Shuster and Farley (0.43 mm, which is within the 0.25-0.5 mm grain size fraction that we analysed).

We further assume that a fire of duration  $t_f$  (s) takes place at the beginning of each 250-yr time step.  $L_{ij}$  is then proportional to the sum of the quantity  $Dt_f$  for all time steps from  $j$  to  $m$ , which is:

$$\sum_{k=j}^m D_{i,k} t_f$$

$L_{ij}$  is computed using Equation (4) of Fechtig and Kalbitzer (1966). As for  $^{10}\text{Be}$ , the resulting  $^{21}\text{Ne}$  concentrations at the end of the model run are averaged for all particles in each 10-cm depth increment.

Given measured or assumed values for  $D_0$  and  $E_a$  for  $^{21}\text{Ne}$  in quartz and a fire frequency defined by the time step length  $\delta t$ , the adjustable parameters for the diffusive loss part of the model are the inherited  $^{21}\text{Ne}$  concentration  $N_{21,inh}$ , the surface temperature during a fire  $T_0$ , the heating attenuation constant  $\Lambda_T$ , and the fire duration  $t_f$ . As noted in the main text, many of these parameters are not independent: the controlling parameter for  $^{21}\text{Ne}$  loss is the product  $Dt_f$ , which can be varied by changing the fire temperature profile, the fire duration, and/or the fire frequency. For the example model runs shown in Figure 3 in the main text we used a time step of 250 yr (which approximates the long-term fire frequency in Long et al., 2018) and heating parameters  $T_0 = 450^\circ \text{C}$ ,  $\Lambda_T = 6 \text{ cm}$ , and  $t_f = 60$  minutes (which approximate a typical wildfire heating scenario in Richter et al., 2022 or similar work).  $N_{21,inh}$  is lithology-specific and has not been estimated for the source bedrock at our study site, so is unconstrained. We assumed  $N_{21,inh} = 2 \text{ Matoms/g}$ , which is a reasonable value for quartz with typical U/Th concentrations and a Cenozoic cooling age, that approximately fits the data. In future work, this parameter could be measured in shielded samples of the source bedrock.

A MATLAB implementation of the model is provided as a supplemental file.

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