Spatial variability of Quaternary denudation rates across a volcanic ocean island (Santo Antão, Cape Verde) from cosmogenic ^3He

Camille Litty, Julien Charreau, Pierre-henri Blard, Raphaël Pik, Sébastien Nomade

► To cite this version:

Camille Litty, Julien Charreau, Pierre-henri Blard, Raphaël Pik, Sébastien Nomade. Spatial variability of Quaternary denudation rates across a volcanic ocean island (Santo Antão, Cape Verde) from cosmogenic ^3He . *Geomorphology*, 2021, 375, pp.107557. 10.1016/j.geomorph.2020.107557. hal-03089201

HAL Id: hal-03089201

<https://cnrs.hal.science/hal-03089201>

Submitted on 28 Dec 2020

HAL is a multi-disciplinary open access archive for the deposit and dissemination of scientific research documents, whether they are published or not. The documents may come from teaching and research institutions in France or abroad, or from public or private research centers.

L'archive ouverte pluridisciplinaire **HAL**, est destinée au dépôt et à la diffusion de documents scientifiques de niveau recherche, publiés ou non, émanant des établissements d'enseignement et de recherche français ou étrangers, des laboratoires publics ou privés.

Spatial variability of Quaternary denudation rates across a volcanic ocean island (Santo Antão, Cape Verde) from cosmogenic ^3He

Camille Litty^{1,2}, Julien Charreau², Pierre-Henri Blard², Raphael Pik², Sébastien Nomade³

¹Université Grenoble Alpes, CNRS, IRD, IFSTTAR, ISTerre, 38000 Grenoble, France

²Centre de Recherches Pétrologiques et Géochimiques, CRPG Nancy, France

³Université Versailles Saint-Quentin et Paris-Saclay, CEA Saclay, CNRS UMR 8212 Laboratoire des Sciences du Climat et de l'Environnement, Gif-sur-Yvette, France

Corresponding authors : Dr. Camille Litty, camille.litty@univ-grenoble-alpes.fr

Abstract

Denudation of volcanic ocean islands creates remarkable landscapes and contributes to Earth's carbon cycle, since the chemical alteration of basalts is a CO_2 sink. Because many volcanic islands have large climate gradients and relatively low variations in lithology and tectonic history, they represent excellent natural laboratories for studying climatic effects on landscape evolution. However, little is known about the control of denudation rates in ocean islands and the respective influences of climatic gradients and morphological parameters. Here, we present new measurements of long-term denudation rates from Santo Antão Island, Cape Verde (17°N). In this 779 km^2 island, mean annual precipitation ranges from 40 mm.yr^{-1} in the southwest to 1100 mm.yr^{-1} in the north. To constrain the spatial distribution of denudation rates, we measured the cosmogenic ^3He concentration in river-transported pyroxene grains from 23 river bedload samples. We obtained basin wide denudation rates ranging from 2.7 ± 0.1 to $57.5 \pm 0.3 \text{ m/Ma}$. The denudation rates display a significant spatial variability, with the highest rates in catchments located in the northeast side of the island where modern precipitation are the highest and low denudation rates in the southern and western dry basins. Our study shows that precipitation is the main control on denudation and landscape development of the Santo Antão volcanic island. This study provides for the first time the spatial distribution of denudation rates across a volcanic island located in a tropical zone.

1. Introduction

Climate is thought to be one of the main parameters controlling denudation rates (e.g., Portenga and Bierman, 2011), which in turn impact many processes such as the building and evolution of mountain ranges (e.g., Willett, 1999; Whipple, 2009). Understanding the effects of climate on denudation rates is therefore thought to be critical, but the modality of its impact remain a matter of long-standing debate (e.g., Langbein and Schumm, 1958; DiBiase and Whipple, 2011). Conversely, denudation may also exert a strong feedback on climate at the geological time scale (e.g., Berner et al., 1983). Silicate weathering on continents consumes CO_2 and releases cations (Ca, Mg, K, Na) to the ocean. In the ocean, Ca and Mg are then combined with CO_3^{2-} and precipitated into bicarbonate, resulting

in a net CO₂ sink (e.g., Berner et al, 1983). The weathering of basaltic lithologies are thus supposed to be an efficient CO₂ sink at the geological time scale (Dessert et al., 2003).

In such context, volcanic islands in tropical areas are noteworthy natural laboratories allowing the investigation of links and interactions between denudation and climatic processes for two main reasons. First, the amplitude of the denudation variations can be compared to climate only. Indeed, most of the volcanic islands often exhibit a strong climatic gradient, while tectonic and lithological forcing are comparatively negligible (Hildenbrand et al., 2008; Salvany et al., 2012; Jefferson et al., 2014). Second, according to geochemical measurements of dissolved load in rivers, the modern weathering of volcanic islands, especially in tropical areas, is especially intense (e.g., Louvat and Allègre, 1997, Jefferson et al., 2014). It may represent up to 35% of the total CO₂ consumed from continental silicate weathering (Gaillardet et al., 1999) and, hence, have a large influence on long-term climate changes (Lenton and Britton, 2006).

In this general context, it is of the utmost importance to better constrain denudation rates (i.e., the sum of physical and chemical erosion) and their spatial distribution across these islands. Several studies have estimated the long-term (10⁵-10⁶ a) denudation rates of several islands by reconstructing the original volcanic surfaces and then by calculating the volumes removed by erosion (Hildenbrand et al., 2008; Salvany et al., 2012; Menéndez et al., 2008; Ferrier et al., 2013; Ricci et al., 2015). This geomorphological approach relies on the postulate that the construction of the volcanoes precedes their destruction by erosion. However, these two processes can be indeed coeval, which can hamper an accurate estimate of the initial pre-erosion surface. Moreover, it is often hazardous to accurately estimate the age of the original surfaces of the volcanoes. The derived denudation rates are therefore tied to large uncertainties and must be rigorously considered as minimum values (Hildenbrand et al., 2008). Other studies have constrained modern denudation rates from direct geochemical measurements of dissolved load in rivers (Louvat and Allègre, 1997; Ferrier et al., 2013). However, these estimates remain only strictly valid over short time scales (10¹-10²a). Therefore, the mechanisms by which volcanic islands erode at the Quaternary time scale (10²-10⁴) remain largely unknown despite being a critical parameter for a comparison with long-term climate change.

Basin wide denudation rates at the Quaternary time scale can be quantified by measuring the concentration of cosmogenic nuclides in river sediment. In-situ cosmogenic nuclides are produced by the flux of secondary cosmic particles in the top few meters of Earth's surface (e.g., Granger et al., 1996). However, denudation rate measurements from cosmogenic nuclides have been mainly based on quartz-rich lithologies through ¹⁰Be analyses (e.g., Granger et al., 1996). On basaltic and andesitic islands, quartz is rare or inexistent, implying that ¹⁰Be cannot usually be used (Blard et al., 2008). Cosmogenic ³He measured in pyroxene and olivine in river bedload sediment represents a good alternative to ¹⁰Be for these mafic terrains (e.g., Puchol et al., 2017). These minerals have a high helium retentivity (Blard and Pik, 2008) and their ³He cosmogenic production rate is relatively well established (Balco et al., 2008). ³He has been widely used for the past 30 years to date lava flows (e.g., Kurz, 1986) or to

reconstruct continental paleoclimates (e.g., Blard et al., 2007). However, only a few studies have used cosmogenic ^3He to estimate denudation rates (Gayer et al., 2008, Ferrier et al., 2013, Puchol et al., 2017). Sarda et al. (1993) were the first to determine denudation rates from cosmogenic ^3He in a volcanic island, but this study was limited to a depth profile at one single location in La Reunion. To quantify Quaternary denudation rates Gayer et al. (2008) and Ferrier et al. (2013) measured the cosmogenic ^3He in river sand in Kaua'i Island (Hawaii), but both studies were limited to a single drainage basin.

In this paper, we analyze the ^3He cosmogenic concentrations in pyroxenes and olivines of 23 sand samples collected in 19 different modern riverbeds across the entire Santo Antão Island (Cape Verde). We hence determine, for the first time across a volcanic island, the near-whole spatial variability of denudation rates averaged over the last 10^3 to 10^5 ka.

2. Geological and climatic settings of the Cape Verde islands and Santo Antão

The Cape Verde archipelago (4033 km²) includes 10 major and several smaller islands located ~600 km west of the African continent (Fig. 1). The development of the islands likely initiated 19 Ma ago with tholeiitic sills that intruded the Jurassic–Cretaceous sea-floor basement (Hayes & Rabinowitz, 1975; Duncan & Jackson, 1977). Since 19 Ma, the volcanic activity has not migrated and has been well distributed across the islands until present (Mitchell et al., 1983; Plesner et al., 2002). The active volcanic activity is now limited to Fogo Island with the last eruption in 1995.

The climate of the Cape Verde archipelago is driven by the respective position of the subtropical high pressure of the North Atlantic (Açores anticyclone) and the Intertropical Convergence Zone (Mannaerts and Gabriels, 2000). The wet season lasts from July to October when the ITCZ, usually centered around the equator, shifts toward 10–20°N and brings 80% of the annual precipitation, whereas during the rest of the year, precipitation is scarce (Mannaerts and Gabriels, 2000).

Santo Antão is the northernmost and second largest island (770 km²) of the archipelago (Fig. 1). Its elevation ranges from 0 to 1979 m (Tope de Coroa), with a mean elevation of 362 m. The relief of the island forms a major orographic barrier that creates a large precipitation gradient with annual precipitation that varies by a factor of 20–30 between the wettest northeast and the driest southwest sides (Fig. 3a). Local rainfall is also strongly dependent on elevation, with some low coastal areas having low rainfall of 0 – 150 mm.yr⁻¹ whereas the mountain areas above 1000 m elevation receive ~800 – 1000 mm.yr⁻¹ (Fig. 3a; Mannaerts and Gabriels, 2000). This precipitation gradient seems to have induced significant topographic heterogeneities across the island: major valleys on the north and northeast side of the island receive regular rain during the wet season and are deeply incised. In contrast, the southern part of the island receives only very irregular rain and the topography exhibits wide smooth seaward slopes preserving volcanic surfaces that are poorly incised (Fig. 3b). In addition, two deep (>1 km) and large (> 5km) circular depressions are located in Tarrafal and the Cha di Morte regions (Fig. 2 and 3). These two depressions together with the deeply entrenched valley in the north provide numerous

outcrops of the lavas that built the island and enable a good reconstruction of its volcanic history (e.g., Plesner et al., 2002; Holm et al., 2005).

On Santo Antão, the oldest lavas reported are ~7.5 Ma old altered picrites and ankaramites of the Dos Bodes formation. They can be observed in a limited area at the bottom of the Cha di Morte depression (Plesner et al., 2002; Holm et al., 2005; Fig. 2). This formation is overlain by the thick (~800 m) 3-1 Ma old Cha di Morte formation (Fig. 2; Plesner et al., 2002). The lava flows of this formation dip toward the coast and point toward sources of magma emission located in the central part of the island. The Dos Bodes and Cha di Morte formations are usually mapped together as “Older Volcanics” by several authors (Fig. 2; Holm et al., 2005). After the emplacement of the Cha di Morte formation, between ~1.4 Ma and 0.3 Ma, other lavas flows were emplaced in other areas of the island, indicating that the recent volcanic activity was less focused and more distributed (Plesner et al., 2002; Holm et al., 2005). The Tope de Coroa volcano experienced a large flank collapse that is not yet dated and yields to the large Tarrafal circular depression. This depression was later refilled by the development of the modern Tope de Coroa stratovolcano ~0.1 Ma ago. Since ~0.3 Ma, the volcanic activity of the island is mainly characterized by explosive magmatism including numerous individual scoria cones, phreatomagmatic to strombolian and several pluri-meter-thick phonolitic to tephri-phonolitic tephra deposits referenced as the Cao Grande formation (Plesner et al., 2002; Eisele et al., 2015). During this recent explosive period, several lava flows are reported in the Tope de Coroa and in the Porto Novo region.

3. Material and methods

3.1 Basin average denudation rates

Assuming isotopic steady state, meaning that the cosmogenic production of ^3He is balanced by the erosional loss (Lal, 1991), approximating (similarly to the spallogenic production) the two muogenic production rates at depth with two different exponential laws, then the basin average denudation rate (cm.yr^{-1}) is given by:

$$\bar{\varepsilon} = \sum_{x=sp, \mu_{sm}, \mu_{fm}} \frac{\overline{P_{x,3}} \Lambda_x}{\rho {}^3\text{He}_c} \quad (1)$$

where the subscripts sp , μ_{sm} and μ_{fm} correspond to the production by spallation, slow muon capture and fast muon processes, respectively. $\overline{P_{x,3}}$ is the basin average *in situ* cosmogenic production rates at the surface ($\text{at.g}^{-1}.\text{a}^{-1}$). ρ and Λ_x are the rock density (g.cm^{-3}) and the attenuation length in the rocks (g.cm^{-2}) of each production pathway x , respectively. ${}^3\text{He}_c$ (at/g) is the mean concentration of the cosmogenic ${}^3\text{He}$ in the river sediment. We use a basalt density of 3 g/cm^3 and assumed constant attenuation lengths of 160 g/cm^2 , 4320 g/cm^2 and 1500 g/cm^2 for spallation, fast muons capture and slow muons processes (e.g. Braucher et al., 2011). The basin average production rates were computed using a cell-by-cell

approach and the SIG tool Basinga (Charreau et al., 2019) with a 30 m digital elevation model of the Santo Antão Island. The production rates of each of the cells of the DEM were scaled using the time dependent Lal-Stone scheme (Stone, 2000), and a reference surface production at sea level and high latitude of $122 \pm 12 \text{ at.g}^{-1}.\text{a}^{-1}$ has been used (world-wide mean of all calibration studies computed with the CREp calculator, Martin et al., 2017). Following DiBiase (2018), we did not apply any topographic shielding correction.

3.2 Determination of cosmogenic ^3He in pyroxenes and olivines

The total ^3He concentration ($^3\text{He}_{\text{tot}}$) in both pyroxenes and olivine is a three-component system (Blard and Farley, 2008) that can be written as follows:

$$^3\text{He}_{\text{tot}} = ^3\text{He}_c + ^3\text{He}_{\text{mag}} + ^3\text{He}_{\text{nuc}} \quad (2)$$

where indices *c*, *mag* and *nuc* refer to the concentration in matrix-sited cosmogenic ^3He , the magmatic ^3He located in fluid and melt inclusions and the nucleogenic ^3He produced by neutron capture on Li nuclei since the eruption of the lavas, respectively.

Similarly, the total ^4He budget is a two-component system consisting of magmatic ^4He inherited in fluid and melt inclusions and matrix-sited radiogenic ^4He mainly produced by the decay of ^{238}U , ^{235}U and ^{232}Th since the eruption:

$$^4\text{He}_{\text{tot}} = ^4\text{He}_{\text{rad}} + ^4\text{He}_{\text{mag}} \quad (3)$$

Therefore, the cosmogenic $^3\text{He}_c$ can be estimated from Eq (2) if the magmatic, the nucleogenic and the total ^3He can be estimated independently with accuracy and precision. The $^3\text{He}_{\text{tot}}$ can be measured by melting the sample. The nucleogenic contribution $^3\text{He}_{\text{nuc}}$ can be calculated using the mean neutron flux, the Li concentration and the eruption age of the minerals (Andrews, 1982). In practice, we used here the major and trace elements and the Li-concentrations of the phenocrysts measured at the SARM in Nancy to compute this nucleogenic component (Table A.1).

Because it can be quite variable from one aliquot to another (e.g.s Puchol et al., 2017), the magmatic $^3\text{He}_{\text{mag}}$ component is difficult to estimate. Magmatic He can be preferentially released and analyzed by in vacuo mineral crushing (Kurz, 1986). The crushing must, however, be sufficiently long to release and purge most of the trapped magmatic He. On the other hand, this crushing should not be too intense nor too long to avoid releasing the matrix sited $^3\text{He}_c$ and/or the $^4\text{He}_{\text{rad}}$ (Blard et al., 2006, 2008), nor contaminate the sample with atmospheric He (Protin et al., 2016).

In the present study, we crushed all the pyroxenes/olivines and analyzed the ^3He and ^4He concentrations extracted during this extraction step (“crush” indice). We fused in a high vacuum furnace additional uncrushed pyroxene/olivine aliquots (above 1400°C , to ensure a complete He release), and measured the ^3He and ^4He concentrations release (“melt” indice). Then, we describe and apply two approaches to estimate the magmatic component and to calculate the ^3He cosmogenic concentration. Estimation of $^3\text{He}_{\text{nuc}}$ is the same in both approaches.

The first approach is derived from the initial method proposed by (Kurz, 1986), and includes the latter improvements by Dunai et al. (2000) and Blard and Farley (2008), to account for the presence of $^4\text{He}_{\text{rad}}$. In this method, the cosmogenic ^3He concentration is calculated as follow:

$$^3\text{He}_c = ^3\text{He}_{\text{tot}} - ^3\text{He}_{\text{mag}} - ^3\text{He}_{\text{nuc}}$$

$$^3\text{He}_c = ^3\text{He}_{\text{tot}} - [^4\text{He}_{\text{mag}} \times (^3\text{He}_{\text{mag}} / ^4\text{He}_{\text{mag}})] - ^3\text{He}_{\text{nuc}}$$

from Eq (3):

$$^3\text{He}_c = ^3\text{He}_{\text{tot}} - [(^4\text{He}_{\text{tot}} - ^4\text{He}_{\text{rad}}) \times (^3\text{He}_{\text{mag}} / ^4\text{He}_{\text{mag}})] - ^3\text{He}_{\text{nuc}}$$

and then:

$$^3\text{He}_c = ^3\text{He}_{\text{melt}} - [(^4\text{He}_{\text{melt}} - ^4\text{He}_{\text{rad}}) \times (^3\text{He}/^4\text{He})_{\text{crush}}] - ^3\text{He}_{\text{nuc}} \quad (4)$$

The $(^3\text{He}/^4\text{He})_{\text{crush}}$ is the isotopic ratio measured during the initial crushing of the pyroxenes and olivines, which is considered as representative of the magmatic isotopic ratio.

Because pedogenesis and sediment transport induces break up of phenocrysts, we consider that the pyroxenes and olivines sampled in the riverbed loads do not bear any implanted $^4\text{He}_{\text{rad}}$ from the (U-Th) richer lava (Blard and Farley, 2008). This is justified because the implanted component is only present in the 20 μm outer rim (e.g., Dunai et al., 2000). Consequently, we computed the $^4\text{He}_{\text{rad}}$ with the U-Th-Sm concentrations measured in the sole phenocrysts (Tables 1 and A.2). To estimate the mean ages of the lava flow, we used the geological map of List et al. (2007), and weighted each eruption age with the respective contributive size of each studied basin. Additionally, if the measured ^4He concentration during melting is lower than the calculated $^4\text{He}_{\text{rad}}$, then we assume that the $^4\text{He}_{\text{mag}}$ is null in these aliquots and we only apply a correction for $^3\text{He}_{\text{nuc}}$.

The second approach assumes that the in vacuo crushing step was efficient enough to release all the magmatic ^3He from the aliquots and that the measured concentrations in the crushed aliquots are representative of those of the melted aliquot. In this case, Eq. (2) can be rewritten as follows:

$$^3\text{He}_c = ^3\text{He}_{\text{melt}} - ^3\text{He}_{\text{crush}} - ^3\text{He}_{\text{nuc}} \quad (5)$$

with the indices *melt* and *crush* referring to the ^3He measured after melting and crushing experiments of the sample. The calculated cosmogenic ^3He concentrations from both approaches, Eqs. (4) and (5) are then used to estimate the average denudation rates in the different sampled river catchments on Santo Antão Island applying Eq. (1) (e.g., Granger et al., 1996).

3.3 Sampling, treatment and analyses

Twenty-three samples were collected at the mouth of the main rivers all around the island (Fig. 3, Table A.1). Approximately 3 kg of river bedload sediment was collected from each sampling site. Bulk sands were sieved under flowing water into the 0.15–0.5 mm granulometric fraction. A hand magnet was first used to remove a large fraction of the magnetic sediment. Then the pyroxene and olivine grains were concentrated from the sand by two magnetic separations using a Frantz device. Nearly pure pyroxene and olivine grains were then obtained by gravimetric separation in di-iodomethane (density $\sim 3.3 \text{ g cm}^{-3}$). After washing in diethyl ether and drying, the grains were cleaned in an ultrasonic

bath of deionized water, and then acetone, for 10 min each, to remove any impurities resulting from dust. After drying, all grains with attached lava fragments were removed and discarded by handpicking under a binocular microscope. We chose to work mainly on pyroxenes because they were more abundant in our samples and thus easier to separate. Nevertheless, when olivines were present in some samples, we also analyzed mixed olivine/pyroxene aliquots. To measure the total ^3He and ^4He concentrations, grain aliquots were wrapped in copper foils and fused at 1350°C for 15 min. High temperature extractions were performed with the full metal induction furnace designed at CRPG (Zimmermann et al., 2012). After extraction, the gas was purified using activated charcoal at -196°C and with hot (400°C and 600°C) and ambient temperature (21°C) titanium sponge getters. A complete description of the furnace and the purification line can be found in Zimmermann et al. (2012). Helium was analyzed on a GV HELIX Split Flight Tube dual collector mass spectrometer. The instrument sensitivity was determined using the HESJ standard (Matsuda et al., 2002) whose certified $^3\text{He}/^4\text{He}$ ratio is 20.63 ± 0.10 Ra ($\text{Ra} = 1.384 \times 10^{-6}$ being the atmospheric $^3\text{He}/^4\text{He}$ ratio). To measure the He_{mag} concentration, in-vacuo crushing extractions were performed on additional aliquots with a soft iron slug activated by external solenoids (5 min at 100 strokes/min). After a first measurement, to optimize the release of all the He_{mag} , second crushing extractions of the same aliquot were performed. In the first approach, the mean value of the two crushing experiments was used to calculate the $(^3\text{He}/^4\text{He})_{\text{crush}}$ ratios (i.e., the magmatic $^3\text{He}/^4\text{He}$ ratio). In the second approach, the sum of the two crushing experiments has been used to compute the $^3\text{He}_c$ as this approach assumes that the crushing extracts the whole magmatic He. Finally, to test the reproducibility of the methods, nine samples have been replicated and analyzed after high temperature extraction. For these replicates, magmatic corrections were performed using the same He results from in-vacuo crushing extraction.

4. Results

Information about samples and catchment metrics are given in supplementary Table A.1. Annual precipitation rates have been estimated from the TRMM model with 20% of uncertainties and gradient information have been extracted from a 30 m resolution DEM (Table A.1). Helium results of the pyroxenes/olivines crushing are presented in Table A.1. Major and trace element concentrations used to calculate the $^4\text{He}_{\text{rad}}$ and $^3\text{He}_{\text{nuc}}$ components are given in supplementary Table A.2. Results of high temperature extractions and calculated $^3\text{He}_c$ concentration and denudation rates are presented in Table 1.

Mean ^4He and ^3He blanks during heating extraction were $(7.3 \pm 0.3) \times 10^{-15}$ mol and $(1.24 \pm 0.2) \times 10^{-19}$ mol, respectively. Repeated measures after sample heating extraction always yielded He concentrations in the blank range, indicating that the extraction was complete after the first heating step. Mean ^4He and ^3He sample values during heating extraction before blank correction were $(1.10 \pm 0.03) \times 10^{-12}$ mol and $(4.2 \pm 0.2) \times 10^{-18}$ mol, respectively, one to three orders of magnitude higher than the

measured blank concentrations. This implies that blanks do not represent a significant source of uncertainty. Mean ^4He and ^3He blanks measured from the crushing device (with the slug moving) were $(8.6 \pm 0.4) \times 10^{-15}$ mol and $(1.4 \pm 0.1) \times 10^{-19}$ mol, respectively. Mean ^4He and ^3He sample values for the crushing experiments before blank correction were $(5.8 \pm 1.5) \times 10^{-14}$ mol and $(6.5 \pm 1.7) \times 10^{-19}$ mol, respectively. Blanks therefore represent 10 to 20% of the crushing concentrations. The calculated concentrations have been then corrected with the mean blank values.

This corresponds to total ^4He and ^3He concentrations measured after heating experiments ranging between $(13 \pm 1) \times 10^{10}$ and $(27 \pm 1) \times 10^{11}$ at/g and between $(22 \pm 3) \times 10^5$ and $(29 \pm 1.3) \times 10^6$ at/g, respectively (Table 1). The magmatic ^4He and ^3He concentrations measured after the first in-vacuo crushing extractions range between $(6.3 \pm 0.1) \times 10^9$ and $(2.6 \pm 0.2) \times 10^{11}$ at/g and between $(1.9 \pm 0.1) \times 10^4$ and $(8.8 \pm 0.4) \times 10^6$ at/g, respectively (Table A.1). Corresponding $^3\text{He}/^4\text{He}$ ratios measured by crushing range between 2 ± 14 Ra and 15 ± 10 with a weighted mean of 12.2 ± 0.3 Ra ($\text{Ra} = 1.384 \times 10^{-6}$ being the atmospheric $^3\text{He}/^4\text{He}$ ratio). Second in-vacuo crushing extractions show similar results and, when summed with the first crushing, returned total magmatic values that range between $(6.3 \pm 0.1) \times 10^9$ and $(5.8 \pm 0.2) \times 10^{11}$ at/g and between $(1.9 \pm 0.1) \times 10^4$ and $(1.1 \pm 0.06) \times 10^7$ at/g for ^4He and ^3He , respectively. The isotopic ratios of the two crushing together range between 2 ± 14 Ra and 14.1 ± 0.9 Ra, with a weighted mean of 11.8 ± 0.4 Ra. However, some samples have not been measured for a second in-vacuo crushing extraction.

With the first approach, calculated cosmogenic ^3He concentrations range between $(2.1 \pm 1.4) \times 10^6$ at/g and $(2.0 \pm 0.17) \times 10^7$ at/g (Table 1). Eight aliquots show $^3\text{He}_{\text{mag}} > ^3\text{He}_{\text{melt}}$ therefore yielding null cosmogenic $^3\text{He}_c$ concentrations. The second approach leads to $^3\text{He}_c$ concentrations ranging between $(2.0 \pm 0.3) \times 10^6$ and $(2.3 \pm 0.1) \times 10^7$ at/g. Importantly, the two approaches yield $^3\text{He}_c$ concentrations and denudation rates that are close to the 1:1 line for most of the samples (Fig. 4a and b), indicating that the choice of the correction procedure for the non-cosmogenic ^3He components is not a significant source of uncertainty. Based on these cosmogenic ^3He concentrations, we determined the denudation rates of 23 samples in 19 main catchments all across the Santo Antão Island. Considering the two approaches, and ignoring the few samples that yielded null-cosmogenic ^3He concentrations, denudation rates show a variability across the island, from 2.7 ± 0.1 m/Ma (CV18-01) to 58 ± 4 m/Ma (CV18-20), with a mean denudation rate over the studied catchments of 18.0 ± 5.2 m/Ma. Finally, the nine replicated measurements are compatible within analytical uncertainties (Fig. 5a, b,c). For denudation rates lower than 15m/Ma, the correlation between replicates is robust and is not dependent on the approach (Fig. 5d; $R^2 = 0.76$, p-value = 0.002).

5. Discussion

5.1 Reliability of the calculated cosmogenic ^3He concentrations and derived denudation rates

The reconstruction of the $^3\text{He}_c$ concentrations relies on several assumptions and corrections that must be considered. First, the two approaches give similar results, which suggest that the concentration in cosmogenic ^3He has been accurately quantified. However, when looking in detail, the $^3\text{He}_c$ concentrations obtained with the second approach are lower by ~10 to 20% compared to those of the first approach, for most of the samples (Fig. 4a), leading to ~10 to 20% higher denudation rates (Fig. 4b). The first approach depends on the accuracy of the ^4He radiogenic and the magmatic ($^3\text{He}/^4\text{He}$) estimates. The assumption that the implanted radiogenic ^4He from the (U-Th) richer lava can be neglected is supported by the fact that, if it is considered, computed radiogenic ^4He are larger than the total ^4He measured in almost 90% of the analyzed aliquots. On the other hand, the lower $^3\text{He}_c$ concentrations obtained with the first approach could suggest an underestimate of the $^4\text{He}_{\text{rad}}$. This could result from an underestimate of the eruption age of the bedrock material. Rivers may actually erode more efficiently the older lithologies than the younger ones. Alternatively, older rocks are porphyritic and probably contain more pyroxenes than the younger ones. However, we carefully estimated the mean bedrock ages in each basin, propagating high uncertainties considering the age distributions within a given basin (from 10 to 93%, average = 60%; Table A.1). For that reason, the first approach yields higher uncertainties (mean 1σ error of 20%) than the second one (mean 1σ error of 10%). Second, the lower cosmogenic ^3He concentrations calculated with the first approach compared to the second may also suggest an overestimation of $(^3\text{He}/^4\text{He})_{\text{mag}}$ ratio from the crushing experiments. Too intense crushing could have partially released the $^3\text{He}_c$ trapped in the crystal lattice (Blard et al., 2006, 2008). Moreover, the good reproducibility between the nine replicates suggests that merging hundreds of grains in each aliquot seems sufficient to average the $^3\text{He}_c$ concentration, despite possible spatial variability of the denudation in the watershed (Puchol et al., 2017). Samples yielding denudation rates above ~15 m/Ma display an increase of the inter-aliquot scatter, with up to 57% of differences between the replicates (Fig. 5c). This observation suggests that, in Santo Antão Island, ~15 m/Ma seems to be a threshold value, above which denudation rates can only be measured with lower accuracy and precision. This threshold is directly controlled by the cosmogenic/magmatic ratio, as well as by the amount of radiogenic ^4He . It may thus vary from one geological setting to another and it may represent a significant limitation, particularly in catchments where denudation rates are high ($\gg 10$ m/Ma; Puchol et al., 2017).

Foeken et al. (2012) have calibrated the local ^3He cosmogenic production rate in the nearby Fogo Island to 75.6 ± 7.0 at.g $^{-1}$.yr $^{-1}$ in olivine and 70.1 ± 2.9 at.g $^{-1}$.yr $^{-1}$ in pyroxene. These values are significantly lower than other ^3He production rates found elsewhere in the world and the reference surface production at sea level and high latitude of 122 ± 12 at.g $^{-1}$.yr $^{-1}$ (Martin et al., 2017) that we use in our calculations. Foeken et al. (2012) explained this discrepancy by geomagnetic field strength spatial variability. Considering these local production rates would decrease all our Cape Verde denudation rate values but would not change their relative spatial variations.

In summary, we conclude that we accurately measured and estimated the different He components and that the two approaches provide reliable estimates of the cosmogenic $^3\text{He}_c$ concentration. Because the uncertainties associated to the second approach are lower than those of the first one, and because both approaches produce results that are identical within uncertainties, we restrain the following discussion to the denudation rates derived from the second approach.

5.2 Denudation rates variability in Santo Antão

The studied hydrological basins of Santo Antão are characterized by a broad range of basin-averaged denudation rates, ranging between ~ 3 and ~ 60 m/Ma and therefore varying by a factor of 20 across the island (Fig. 6; Table 1). The basins in the south and west sides of the island display the lower calculated denudation rates ranging from ~ 3 and ~ 20 m/Ma. One sample (CV18-26) located in the south of the island but shows a relatively high denudation rate (26.1 ± 5.4 m/Ma and 36.6 ± 4.4 m/Ma with the first and second approach respectively) and compared to the other southern basins seems to be an outlier that is not representative of this subregion. The northern basins are characterized by a gradient of the denudation rates from the southeast to the northwest. Finally, the region of the Cha di Morte displays the highest denudation rates of the island, with values reaching ~ 60 m/Ma (Fig. 6). These denudation rates inferred from $^3\text{He}_c$ are averaged over characteristic timescales ($T_{\text{int}} = (\Lambda/\rho)/\varepsilon$). In Santo Antão, the integrated timescale of these denudation rates ranges from ~ 10 ka for the fastest eroded basins in the north, to ~ 200 ka for the southern basins, with a mean value of ~ 50 ka for the island (Table 1). However, in the south side of the island, the calculated integration time is in some places higher than the age of the rocks. This suggests that the southern side could exhibit even lower denudation rates than the calculated rates. This variability of long-term denudation rates variability across Santo Antão seem to correlate with the modern precipitation pattern on the island. Basins located in the northeast and receiving relatively high precipitation are characterized by the highest denudation rates. Basins located in the south receiving the lowest annual precipitation rates of the island yield the lowest denudation rates (Fig. 3a and 6). This observation is consistent with the study of Ferrier et al. (2013): by reconstructing the eroded volume in 24 basins across Kaua’I in Hawaii, these authors found that long-term (Ma) denudation rates on the island are correlated with modern annual precipitation. However, given the typical integration time of our cosmogenic ^3He -derived denudation rates of 53 ka, spatial patterns in precipitation rates may have been different in the past, meaning that the denudation rates we have measured could result from superimposed successive climatic settings. In Santo Antão, the trade winds have been stable since at least the last glacial maximum, 20 ka (Rognon et al., 1996). Another study even suggests that trade winds have been present over the last 140 ka (Hooghiemstra, 1989). Such timescales are similar to, or longer than, the integration time of the cosmogenic ^3He -derived denudation rates. This created a stable mean gradient of precipitation that caused one side of the island to erode at higher rates than the other. Moreover, during this time span the volcanic activity that could

have reshaped and reconstructed the surface of the islands was relatively limited (Plesner et al., 2002). Therefore, climatic conditions can explain the spatial variations of the long-term denudation rates across the island.

On the other hand, the topographic patterns also strongly vary across the island and is also in apparent correlation with the variability of denudation rates. Steeply sloping valleys are located in the northeast side of the island, in the Tarrafal and Cha di Morte depressions where denudation rates are high (> 40 m/Ma). The southern side of the island presents relatively low denudation rates, a relatively smooth morphology and low slope angles (Fig. 7). Gayer et al. (2008) also explained variations in cosmogenic ^3He measured in 26 samples across the Waimea basin in Kaua'i (Hawaii) by a non-linear link between denudation rates and hillslope gradient. Yet the topography of the island could be inherited from its volcanic construction (Lahitte et al., 2012) independently from the precipitation pattern. To better explore and separate the correlation existing between denudation rates, precipitation rates and slopes, we performed a principal component analysis (PCA) using XLSTAT software (Jolliffe, 2002). The PCA for the mean annual precipitation, mean and maximum slopes in the catchments resulted in a distance biplot of the first two principal axes shown in Fig. 8. The first and second axis explained 74.6% and 18.6% of the variance, respectively. Samples were grouped in five clusters according to their denudation rates (Fig. 8). The striking feature of this PCA is that basins with low denudation rates (< 10 m/Ma) present both low mean precipitation and low mean and maximum slopes at the same time. Basins with denudation rates between 10 and 40 m/Ma display both relatively high mean precipitation and relatively high slopes in the catchments. The sample CV18-26, which seemed to be an outlier from a visual point of view, is also an outlier here as it does not belong to the right part of the graph (Fig. 8). Past studies have suggested strong coupling and feedbacks between climate, denudation and landscape evolution in active mountain ranges (e.g., Willet, 1999; Whipple, 2009). This study reinforces this idea and moreover suggests that these coupling and feedback processes also exist on volcanic islands, and that these forcing are not necessarily linear. In Santo Antão, the precipitation patterns over time control the potential for physical erosion and weathering (Jefferson et al., 2014). The spatial variability of denudation rates influences the differential evolution of the morphology over the island, initiating a probable positive feedback: high precipitation led to higher denudation rates on the northeast side of the island, allowing for the creation of deeply incised basins with high slope angles, high drainage density and fluvial valleys that penetrate inland to the central part of the island, amplifying the local denudation. In contrast, low precipitation leading to relatively low denudation rates on the south side of the island causes the formation of weakly dissected basins with low slopes, little tributary development and lack of deep valleys. Over the long term, the morphology of the island may thus have enhanced the denudation variability through positive feedbacks with the slopes where high slopes contributing to increase the potential for fluvial erosion. Therefore, topography on the island seems to be controlled by denudation rates rather than the contrary, at least at the initial stage of the island history. Moreover, climatic conditions seem to be the predominant process that has shaped the morphology of Santo Antão.

However, the highest denudation rates of the island (> 40 m/Ma) are correlated with high precipitation rates, but are not necessarily associated with the highest slopes, suggesting that other parameters, such as geological properties, may also play a role.

Similarly, denudation rates estimated for CV18-19 and CV18-20, which are two of the highest values of the island, but located in the lee side of the island, can also result from other causes. (i) The valley bottom of the Cha di Morte is filled by alluvial conglomeratic deposits that are incised by modern streams. Sediments that have been sampled can possibly be recycled material from paleo conglomeratic deposits. They therefore record paleo-denudation rates that can differ from the modern ones. Additionally, these conglomerates are also probably characterized by a higher erodibility than pristine lava flow basalts. (ii) The Cha di Morte depression has been interpreted as being close to the source of the older volcanic magma and numerous dike and sills are intruding into the Dos Bodes and Cha di Morte formations (Holm et al. 2005). Dikes can significantly influence basin dissection by compartmentalizing the volcanic structure into individual blocks and increase water infiltration (Hildenbrand et al., 2005). Thus, in the Cha di Morte depression, the relative high denudation rates can also possibly be caused by an increase of the chemical alteration or be associated with physical erosion through mass-wasting of individual blocks.

5.3 Denudation rates in other volcanic islands in the world

In Cape Verde, our denudation rates results obtained for Santo Antão, are ranging between ~ 3 and 60 m/Ma and are comparable to those calculated on other volcanic islands characterized by different climatic patterns. In Kaua’I Island, Hawaii, Ferrier et al (2013) computed denudation rates based on eroded volume that range between 3 and 112 m/Ma. For the same island, Gayer et al (2008) proposed a denudation rate of 56 m/Ma from cosmogenic ^3He measured in 26 samples across the Waimea basin. In Gran Canaria, Canary Islands, denudation rates based on eroded volume range between 0.1 and 240 m/Ma (Menéndez et al., 2008). One extreme value reaches 975 m/Ma, however, it has been measured in a small basin of 2.3 km². We decided to discard this value as well as 14 other denudation rates that have been computed in basins smaller than 5 km². This avoids most of the outliers that seem to be related to local effects related to small eroded volumes in modest catchments. Finally, in Tahiti-Nui, in French Polynesia, Hildebrand et al. (2008) calculated denudation rates between 72 and 2700 m/Ma. Results from the different volcanic islands show significant differences in the range and absolute value of the observed denudation rates. Despite the different techniques used to infer denudation rates, previous studies and our study can be directly compared as all techniques measure both the mechanical and chemical erosion (Kirchner et al., 2006; Ferrier et al., 2013). Remarkably, we observe a logarithmic correlation between the Quaternary denudation rates and the mean annual precipitation rates when considering the data of the four islands ($R^2 = 0.63$, $p\text{-value} < 0.001$; Fig. 9). Although these different islands have variable volcanic morphologies and lithology, resulting from different geodynamic contexts, climatic conditions seem to be a predominant control on the denudation rates, and therefore,

on landscape evolution not only in Santo Antão but also in other volcanic islands in general. Our results demonstrate the strong coupling through which climate influences the variability of denudation and show how climate acts to shape the landscape of volcanic islands.

6. Conclusion

We accurately measured long-term (over 10 ka to 200 ka timescale) denudation rates from cosmogenic ^3He in pyroxene from 23 river bedload samples in Santo Antão Island. The calculation of the long-term denudation rates allows the reconstruction of a volcanic island landscape and landform history in response to a contrasted climate. Exposed to the trade winds and associated relatively heavy rainfall, the northeast part of the island exhibits high denudation rates and deeply incised basins. The southern part of the island that currently receives low precipitation displays low denudation rates and gently southward dipping slopes. Accordingly, the denudation rates and the morphological evolution of the Santo Antão Island is probably primarily controlled by climatic processes. After comparison with previously published denudation rates from other islands, our data shows that the large climatic gradients existing on volcanic ocean islands have a predominant role on their landscape evolution. Our data represent additional inputs for construction of the global database of long-term denudation rates affecting volcanoes within tropical settings. The development of such database is key for better quantifying chemical and mechanical erosion processes in tropical environments.

Acknowledgments

The manuscript has been read by all authors, who have approved the content and agree to its publication. This study was supported by the INSU-CNRS SYSTER. We thank Emmanuel Davy for the mineralogical separation of the samples and Laurent Zimmerman for his assistance and help during He measurements.

References

- Andrews, J. N., and R. L. F. Kay. 1982. Natural production of tritium in permeable rocks." *Nature* 298 (5872), 361-363.
- Balco, G., Stone, J.O., Lifton, N.A., and Dunai, T.J., 2008, A complete and easily accessible means of calculating surface exposure ages or erosion rates from ^{10}Be and ^{26}Al measurements: *Quaternary Geochronology* 3 (3), 174–195, doi:10.1016/j.quageo.2007.12.001.

467 Berner, R.A., Lasaga, A.C., and Garrels, R.M., 1983, The carbonate-silicate geochemical cycle and its effect on
468 atmospheric carbon dioxide over the last 100 million years: *American Journal of Science* 283, 641–683, doi: 10
469 .2475 /ajs .283 .7 .641

470 Blard, P.-H., Pik, R., Lavé, J., Bourlès, D., Burnard, P.G., Yokochi, R., Marty, B., Trusdell, F., 2006. Cosmogenic
471 ³He production rates revisited from evidences of grain size dependent release of matrix sited helium. *Earth Planet.*
472 *Science Letter* 247, 222-234.

473 Blard, P. H., Lavé, J., Pik, R., Wagnon, P., Bourlès, D., 2007. Persistence of full glacial conditions in the central
474 Pacific until 15,000 years ago. *Nature* 449 (7162), 591.

475 Blard, P. H., Farley, K. A., 2008. The influence of radiogenic ⁴He on cosmogenic ³He determinations in volcanic
476 olivine and pyroxene. *Earth and Planetary Science Letters* 276 (1-2), 20-29.

477 Blard, P.-H., & Pik, R., 2008. An alternative isochron method for measuring cosmogenic ³He in lava flows *Chem.*
478 *Geol.* 251, 20-32

479 Blard, P.-H., Puchol, N., Farley, K.A., 2008. Constraints on the loss of matrix-sited helium during vacuum crushing
480 of mafic phenocrysts. *Geochim. Cosmochim. Acta*, 72, 3788-3803.

481 Blard, P. H., Bourles, D., Pik, R., & Lavé, J., 2008. In situ cosmogenic ¹⁰Be in olivines and pyroxenes.
482 *Quaternary Geochronology* 3(3), 196-205.
483

484 Charreau, J., Blard, P. H., Zumaque, J., Martin, L. C., Delobel, T., & Szafran, L., 2019. Basinga: A cell- by- cell
485 GIS toolbox for computing basin average scaling factors, cosmogenic production rates and denudation rates. *Earth*
486 *Surface Processes and Landforms* 44 (12), 2349-2365.

487 Dessert, C., Dupré, B., Gaillardet, J., François, L. M., & Allegre, C. J., 2003. Basalt weathering laws and the
488 impact of basalt weathering on the global carbon cycle. *Chemical Geology* 202 (3-4), 257-273.

489 DiBiase, R. A., & Whipple, K. X., 2011. The influence of erosion thresholds and runoff variability on the
490 relationships among topography, climate, and erosion rate. *Journal of Geophysical Research: Earth Surface*
491 116(F4).

492 DiBiase, R. A., 2018. Increasing vertical attenuation length of cosmogenic nuclide production on steep slopes
493 negates topographic shielding corrections for catchment erosion rates. *Earth Surface Dynamics* 6(4).

494 Dunai, T., 2000. Scaling factors for production rates of in situ produced cosmogenic nuclides: A critical
495 reevaluation: *Earth and Planetary Science Letters* 176 (1), 157–169, doi:10.1016/S0012-821X(99)00310-6.

496 Duncan, R. A. & Jackson, E. D., 1977. Geochronology of basaltic rocks recovered by DSDP Leg 41, Eastern
497 Atlantic Ocean. In: Lancelot, Y., Seibold, E., et al. (eds) *Initial Report of the Deep Sea Drilling Project*, 41,
498 Washington, DC: US Government Printing Office, 1113–1118.

499 Ferrier, K. L., Perron, J. T., Mukhopadhyay, S., Rosener, M., Stock, J. D., Huppert, K. L., & Slosberg, M., 2013.
500 Covariation of climate and long-term erosion rates across a steep rainfall gradient on the Hawaiian island of Kaua
501 i. Bulletin 125 (7-8), 1146-1163.

502 Foeken, J. P., Stuart, F. M., & Mark, D. F., 2012. Long-term low latitude cosmogenic ³He production rate
503 determined from a 126 ka basalt from Fogo, Cape Verdes. Earth and Planetary Science Letters 359, 14-25.

504 Gaillardet J., Dupre B., Louvat P., Allegre C.J., 1999. Global silicate weathering and CO₂ consumption rates
505 deduced from the chemistry of large rivers. Chemical Geology, 159 (1–4), 3-30

506 Gayer E., Mukhopadhyay S., Meade B.J., 2008. Spatial variability of erosion rates inferred from the frequency
507 distribution of cosmogenic ³He in olivines from Hawaiian river sediments. Earth Planet. Sci. Lett. 266, 303-315.

508 Granger, D. E., Kirchner, J. W., & Finkel, R., 1996. Spatially averaged long-term erosion rates measured from in
509 situ-produced cosmogenic nuclides in alluvial sediment. The Journal of Geology 104 (3), 249-257.

510 Hayes, D. E., & Rabinowitz, P. D., 1975. Mesozoic magnetic lineations and the magnetic quiet zone off northwest
511 Africa. Earth and Planetary Science Letters 28 (2), 105-115.

512 Hildenbrand, A., Gillot, P.Y., Marlin, C., 2008. Geomorphological study of long-term erosion on a tropical
513 volcanic ocean island: Tahiti-Nui (French Polynesia), Geomorphology 93 (3–4), 460–481,
514 doi:10.1016/j.geomorph.2007.03.012.

515 Hooghiemstra, H., 1989. Variations of the NW African trade wind regime during the last 140 000 years: changes
516 in pollen flux evidenced by marine sediment records. In Paleoclimatology and paleometeorology: modern and past
517 patterns of global atmospheric transport. Springer, Dordrecht. 733-770.

518 Holm, P. M., Wilson, J. R., Christensen, B. P., Hansen, L., Hansen, S. L., Hein, K. M., Mortensen, A. K., Pederson,
519 R., Plesner, S., Runge, M. K., 2005. Sampling the Cape Verde mantle plume: evolution of melt compositions on
520 Santo Antão, Cape Verde Islands. Journal of Petrology 47 (1), 145-189.

521 Jefferson, A. J., Ferrier, K. L., Perron, J. T., Ramalho, R., 2014. Controls on the hydrological and topographic
522 evolution of shield volcanoes and volcanic ocean islands. The Galapagos: A natural laboratory for the Earth
523 Sciences 204, 185-213.

524 Jolliffe, I.T., 2002. Principal Component Analysis, Series: Springer Series in Statistics, 2nd ed., Springer, NY,
525 XXIX, 487 p. 28 illus. ISBN 978-0-387-95442-4.

526 Kirchner, J. W., Riebe, C. S., Ferrier, K. L., Finkel, R. C., 2006. Cosmogenic nuclide methods for measuring long-
527 term rates of physical erosion and chemical weathering. Journal of Geochemical Exploration 88 (1-3), 296-299.

528 Kurz, M.D., 1986. Cosmogenic helium in a terrestrial igneous rock: Nature 320, 435–439, doi:10.1038/320435a0.

529 Lahitte, P., Samper, A., Quidelleur, X., 2012. DEM-based reconstruction of southern Basse-Terre volcanoes
530 (Guadeloupe archipelago, FWI): Contribution to the Lesser Antilles Arc construction rates and magma production.
531 Geomorphology 136 (1), 148-164.

532 Lal, D., 1991. Cosmic ray labeling of erosion surfaces: in situ nuclide production rates and erosion models. *Earth*
533 *and Planetary Science Letters* 104 (2-4), 424-439.

534 Langbein, W. B., & Schumm, S. A., 1958. Yield of sediment in relation to mean annual precipitation. *Eos,*
535 *Transactions American Geophysical Union* 39 (6), 1076-1084.

536 Lenton, T.M., Britton, C., 2006. Enhanced carbonate and silicate weathering accelerates recovery from fossil fuel
537 CO₂ perturbations. *Global Biogeochemical Cycles* 20 (3), GB3009.

538 List, F. K., Klitzsch, E., Kusserow, H., Munier, K., Munier, C., Levenhagen, J., 2007. Cape Verde Islands: Santo
539 Antão, São Vicente, Sal, Fogo, Thematic Maps, Geology. Amt Für Geoinformationsweser der Bundeswehr
540 (AGeoBW)/Geoinformation Office of the German Armed Forces.

541 Louvat, P., & Allègre, C. J., 1997. Present denudation rates on the island of Reunion determined by river
542 geochemistry: basalt weathering and mass budget between chemical and mechanical erosions. *Geochimica et*
543 *Cosmochimica Acta* 61 (17), 3645-3669.

544 Mannaerts, C. M., & Gabriels, D., 2000. Rainfall erosivity in Cape Verde. *Soil and Tillage Research* 55 (3-4), 207-
545 212.

546 Matsuda, J., Matsumoto, T., Sumino, H., Nagao, K., Yamamoto, J., Miura, Y., Kaneoka, I., Takahata, N., Sano,
547 Y., 2002. The ³He/⁴He ratio of the new internal He Standard of Japan (HESJ). *Geochemical Journal* 36 (2), 191-
548 195.

549 Menéndez, I., Silva, P. G., Martín-Betancor, M., Pérez-Torrado, F. J., Guillou, H., Scaillet, S., 2008. Fluvial
550 dissection, isostatic uplift, and geomorphological evolution of volcanic islands (Gran Canaria, Canary Islands,
551 Spain). *Geomorphology* 102 (1), 189-203.

552 Mitchell, J. G., Le Bas, M. J., Zielonka, J., Furnes, H., 1983. On dating the magmatism of Maio, Cape Verde
553 islands. *Earth and Planetary Science Letters* 64 (1), 61-76.

554 Plesner, S., Holm, P. M., Wilson, J. R., 2002. ⁴⁰Ar/³⁹Ar geochronology of Santo Antao, Cape Verde Islands.
555 *Journal of Volcanology and Geothermal Research* 120, 103–121.

556 Portenga, E. W., & Bierman, P. R., 2011. Understanding Earth's eroding surface with ¹⁰Be. *GSA today* 21 (8),
557 4-10.

558 Protin, M., Blard, P. H., Marrocchi, Y., Mathon, F., 2016. Irreversible adsorption of atmospheric helium on olivine:
559 A lobster pot analogy. *Geochimica et Cosmochimica Acta* 179, 76-88.

560 Puchol, N., Blard, P. H., Pik, R., Tibari, B., Lavé, J., 2017. Variability of magmatic and cosmogenic ³He in
561 Ethiopian river sands of detrital pyroxenes: Impact on denudation rate determinations. *Chemical Geology* 448, 13-
562 25.

563 Rognon, P., Coudé-Gaussen, G., Revel, M., Grousset, F. E., Pedemay, P., 1996. Holocene Saharan dust deposition
564 on the Cape Verde Islands: sedimentological and Nd-Sr isotopic evidence. *Sedimentology* 43 (2), 359-366.

- Ricci, J., Lahitte, P., Quidelleur, X., 2015. Construction and destruction rates of volcanoes within tropical environment: Examples from the Basse-Terre Island (Guadeloupe, Lesser Antilles). *Geomorphology* 228, 597-607.
- Salvany, T., Lahitte, P., Nativel, P., Gillot, P. Y., 2012. Geomorphic evolution of the Piton des Neiges volcano (Réunion Island, Indian Ocean): competition between volcanic construction and erosion since 1.4 Ma. *Geomorphology* 136 (1), 132-147.
- Sarda, P., Staudacher, T., Allègre, C. J., Lecomte, A., 1993. Cosmogenic neon and helium at Réunion: measurement of erosion rate. *Earth and Planetary Science Letters* 119 (3), 405-417.
- Willett, S. D., 1999. Orogeny and orography: The effects of erosion on the structure of mountain belts. *Journal of Geophysical Research: Solid Earth* 104 (B12), 28957-28981.
- Whipple, K.X., 2009, The influence of climate on the tectonic evolution of mountain belts. *Nature Geoscience* 2, 97-104, doi: 10.1038/ngeo413.
- Zimmermann, L., Blard, P. H., Burnard, P., Medynski, S., Pik, R., Puchol, N., 2012. A new single vacuum furnace design for cosmogenic ^3He dating. *Geostandards and Geoanalytical Research* 36 (2), 121-129.

Figure captions

Figure 1. Overview of the Cape Verde islands and ASTER satellite images of Santo Antão Island.

Figure 2. Simplified geological map of Santo Antão modified from List et al. (2007).

Fig 3. a. Precipitation and elevation maps over Santo Antão Island from TRMM. The black arrows show the main direction of the wind. b. Photograph of the dry region of Tarrafal on the southwest side of the island. c. Photograph of the wetter valley of the Ribeira on the northeast side of the island.

Fig 4: a. $^3\text{He}_c$ calculated with the first approach (Eq. 4) versus the $^3\text{He}_c$ calculated with the second approach (Eq. 5). b. Denudation rates calculation from the first approach versus the denudation rates calculation from the second approach. The grey band represents the single prediction band of the one standard deviation of the correlation trend in red.

Figure 5: Comparison of the replicate measures of the denudation rates. a. Comparison of the replicates of the denudation rates calculated with the first approach. b. Comparison of the replicates of the denudation rates calculated with the second approach. c. Comparison of the replicates of the denudation rates calculated with the two approaches together. D. Zoom of the Fig. 5c showing the best reproducibility of the measurements for denudation rates lower than ~15 m/Ma.

Figure 6: Spatial variation of the denudation rates on Santo Antão Island.

Figure 7: Map of the slope distribution on Santo Antão and hillshade map on the left top side. The north and northeast side of the island as well show high slopes corresponding to the dissected valleys. The edges of the Cha di Morte also show high slope angles, whereas the south of the island shows relatively low slope angles corresponding to a smooth morphology.

Figure 8: Log-ratio transformed compositional biplots derived from principle component analysis. The length of the black rays is proportional to the amount of variation explained by the corresponding controlling factors. Samples taken from different basins are clustered according to their calculated denudation rates with the second approach.

Figure 9: Denudation rates measured on different volcanic islands in the world using logarithmic scales. The square symbols correspond to computed denudation rates based on eroded volume, whereas the triangles correspond to denudation rates based on ^3He cosmogenic concentration.

Table 1

Sample name	Phenocryst	Fraction	Aliquot mass	³ He(melt)	⁴ He(melt)	1st approach					2nd approach			
						P ₄ (radiogenic)	⁴ He (radiogenic)	P ₃ (nucleogenic)	³ He (nucleogenic)	³ He (cosmogenic)	Denudation rate	³ He (cosmogenic)	Denudation rate	Integration time
		(μm)	(g)	(10 ⁶ at/g)	(10 ¹² at/g)	(10 ⁶ at/g/yr)	(10 ¹² at/g)	(10 ⁻² at/g/yr)	(10 ⁴ at/g)	(10 ⁶ at/g)	(m/Ma)	(10 ⁶ at/g)	(m/Ma)	(ka)
						with U, Th and Sm in lava = 0								
CV18-01	Px	140-500	0.2331	21.7 ± 1.00	1.67 ± 0.05	1.73	1.17 ± 0.39	1.57	1.65 ± 0.55	13.3 ± 1.29	4.67 ± 0.46	19.0 ± 1.02	3.30 ± 0.2	164
CV18-01	Px	140-500	0.1765	25.3 ± 0.06	1.95 ± 0.04	1.73	1.17 ± 0.39	1.57	1.65 ± 0.55	12.1 ± 0.85	5.11 ± 0.39	22.6 ± 0.63	2.70 ± 0.09	195
CV18-02	Px	140-500	0.2569	17.7 ± 0.08	5.22 ± 0.14	2.75	2.18 ± 0.22	0.08	0.08 ± 0.01			13.0 ± 0.88	7.40 ± 0.66	72
CV18-03	Px	140-500	0.2404	17.7 ± 0.08	1.46 ± 0.04	2.25	1.49 ± 0.49	2.74	2.88 ± 0.96	17.7 ± 1.16	4.50 ± 0.31	15.4 ± 0.85	5.20 ± 0.30	103
CV18-03	Px	140-500	0.3403	16.6 ± 0.07	1.21 ± 0.04	2.25	1.49 ± 0.49	2.74	2.88 ± 0.96	16.5 ± 1.10	4.81 ± 0.33	14.3 ± 0.77	5.60 ± 0.32	96
CV18-04	Px	140-500	0.2558	29.0 ± 1.28	2.00 ± 0.06	1.38	0.75 ± 0.25	1.21	1.27 ± 0.42	7.91 ± 1.61	12.7 ± 2.61	18.2 ± 1.37	5.50 ± 0.43	96
CV18-04	Px	140-500	0.1998	30.8 ± 1.36	2.18 ± 0.07	1.38	0.75 ± 0.25	1.21	1.27 ± 0.42	6.68 ± 1.82	15.1 ± 4.12	19.9 ± 1.46	5.00 ± 0.38	106
CV18-05	Px	140-500	0.2374	9.48 ± 0.53	1.28 ± 0.04	2.15	2.91 ± 1.78	1.28	2.30 ± 1.41	9.46 ± 1.36	10.2 ± 1.47	6.34 ± 0.56	15.2 ± 1.38	35
CV18-05	Px	140-500	0.2869	11.0 ± 0.54	1.41 ± 0.04	2.15	2.91 ± 1.78	1.28	2.30 ± 1.41	10.9 ± 1.39	8.80 ± 1.13	7.84 ± 0.58	12.3 ± 0.95	43
CV18-06	Px	75-500	0.2594	8.99 ± 0.49	1.26 ± 0.04	1.55	2.30 ± 2.04	0.80	1.81 ± 1.60	8.98 ± 1.22	9.06 ± 1.24	3.38 ± 0.57	24.0 ± 4.06	22
CV18-06	Px	75-500	0.2546	13.6 ± 0.34	2.39 ± 0.05	1.55	2.30 ± 2.04	0.80	1.81 ± 1.60	12.0 ± 0.96	6.77 ± 0.56	7.94 ± 0.44	10.2 ± 0.61	52
CV18-07	Px	75-500	0.2846	15.7 ± 0.72	1.26 ± 0.04	1.15	2.83 ± 1.64	1.17	5.55 ± 3.22	15.5 ± 1.98	6.41 ± 0.83	9.49 ± 0.79	10.4 ± 0.90	51
CV18-07	Px	75-500	0.2764	20.2 ± 0.45	1.54 ± 0.03	1.15	2.83 ± 1.64	1.17	5.55 ± 3.22	20.2 ± 1.76	4.91 ± 0.44	14.2 ± 0.56	7.00 ± 0.31	77
CV18-08	Px	140-500	0.2884	8.90 ± 0.48	2.01 ± 0.06	2.41	6.10 ± 5.34	1.79	7.17 ± 6.28	8.83 ± 3.34	10.4 ± 3.94	3.87 ± 0.58	23.7 ± 3.58	23
CV18-08	Px	140-500	0.2852	9.18 ± 0.26	1.93 ± 0.04	2.41	6.10 ± 5.34	1.79	7.17 ± 6.28	9.11 ± 3.27	10.1 ± 3.62	4.16 ± 0.41	22.1 ± 2.21	24
CV18-09	Px	140-500	0.2846	3.78 ± 0.29	1.40 ± 0.04	1.90	1.17 ± 0.39	1.29	1.36 ± 0.45			2.76 ± 0.36	33.3 ± 4.41	16
CV18-10	Px	140-500	0.3100	6.25 ± 0.37	1.45 ± 0.04	2.43	3.32 ± 3.07	1.88	4.06 ± 3.76	6.21 ± 1.76	14.8 ± 4.23	4.99 ± 0.44	18.4 ± 1.69	29
CV18-11	Px	140-500	0.2705	2.91 ± 0.27	1.46 ± 0.04	2.44	4.55 ± 3.37	1.21	3.28 ± 2.43	2.88 ± 2.26	31.9 ± 25.11	2.03 ± 0.29	45.2 ± 6.65	12
CV18-11	Px	140-500	0.3562	5.48 ± 0.16	2.55 ± 0.05	2.44	4.55 ± 3.37	1.21	3.28 ± 2.43	5.45 ± 1.87	16.9 ± 5.81	4.61 ± 0.21	20.0 ± 0.98	27
CV18-12	Px/oliv	140-500	0.2466	3.01 ± 0.31	1.51 ± 0.04	2.48	4.18 ± 3.64	1.20	2.77 ± 2.41	2.98 ± 1.92	32.1 ± 20.1	2.96 ± 0.33	32.2 ± 3.68	17
CV 18-12	Px/oliv	140-500	0.2837	2.61 ± 0.19	2.08 ± 0.04	2.48	4.18 ± 3.64	1.20	2.77 ± 2.41	2.58 ± 1.66	37.1 ± 23.8	2.56 ± 0.17	37.4 ± 2.57	14
CV18-13	Px	140-500	0.2459	4.09 ± 0.34	1.33 ± 0.04	1.56	3.37 ± 1.95	1.61	7.63 ± 4.42	4.01 ± 2.64	19.1 ± 12.6	3.59 ± 0.37	21.4 ± 2.24	25
CV18-14	Px	140-500	0.2991	4.23 ± 0.29	0.94 ± 0.03	4.36	8.37 ± 7.28	1.42	3.26 ± 2.84	4.20 ± 3.82	16.6 ± 15.1	3.29 ± 0.37	21.3 ± 2.45	25
CV18-17	Px	140-500	0.3099	4.44 ± 0.29	5.60 ± 0.15	12.5	33.6 ± 29.0	2.86	8.29 ± 7.14	4.36 ± 13.0	21.1 ± 62.8	3.11 ± 0.34	29.5 ± 3.25	18
CV18-18	Px	140-500	0.1818	10.8 ± 0.63	2.70 ± 0.08	4.06	4.58 ± 3.16	1.58	2.29 ± 1.58	10.8 ± 1.96	6.11 ± 1.11	10.1 ± 0.64	6.51 ± 0.44	82
CV18-19	Px/oliv	140-500	0.3134	2.88 ± 0.25	1.57 ± 0.04	0.91	1.08 ± 0.47	0.42	0.87 ± 0.38			2.63 ± 0.28	39.9 ± 4.35	13
CV18-20	Px/oliv	140-500	0.3135	2.53 ± 0.24	3.49 ± 0.09	1.01	1.21 ± 0.53	0.41	0.85 ± 0.37			1.94 ± 0.29	57.5 ± 3.84	9
CV18-25	Px	140-500	0.2216	2.20 ± 0.30	1.12 ± 0.03	0.93	1.57 ± 1.04	1.09	4.92 ± 3.28	2.15 ± 1.43	47.2 ± 31.4	2.08 ± 0.35	47.8 ± 8.16	11
CV18-26	Px/oliv	140-500	0.2916	4.20 ± 0.30	1.42 ± 0.04	1.21	1.47 ± 1.47	0.64	1.29 ± 1.29	4.18 ± 0.85	26.1 ± 5.35	2.98 ± 0.35	36.6 ± 4.37	15
CV18-27	Px/oliv	140-500	0.2925	13.1 ± 0.58	0.13 ± 0.01	13.1	1.25 ± 0.03	3.45	0.36 ± 0.01	13.1 ± 0.80	9.34 ± 0.60	12.6 ± 0.62	9.65 ± 0.51	55
CV18-28	Px/oliv	140-500	0.3783	14.6 ± 0.66	4.80 ± 0.13	4.36	0.93 ± 0.31	2.25	0.67 ± 0.22			12.9 ± 0.68	5.33 ± 0.27	100
CV18-29	Px	140-500	0.2037	13.4 ± 0.68	0.60 ± 0.02	1.26	0.67 ± 0.27	0.74	0.73 ± 0.29	13.3 ± 0.84	6.31 ± 0.42	13.0 ± 0.71	6.46 ± 0.37	83

Figure 1 (Color)

Figure 2 (Color)

Figure 3 (Color)

Figure 4 (Color)

Figure 5 (Color)

a. 1st Approach

b. 2nd Approach

c. The 2 approaches combined

d. zoom of Fig 5c.

Figure 6 (Color)

Figure 7 (Color)

Figure 8 (Color)

Figure 9 (Color)

[Click here to access/download](#)

Supplementary material for online publication only
TableA1.xlsx

[Click here to access/download](#)

Supplementary material for online publication only
TableA2.xlsx

[Click here to access/download](#)

Supplementary material for online publication only
TableA3.xlsx