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Radon signature of CO₂ flux constrains the depth of degassing: Furnas volcano (Azores, Portugal) versus Syabru-Bensi (Nepal Himalayas)

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Substantial terrestrial gas emissions, such as carbon dioxide (CO₂), are associated with active volcanoes and hydrothermal systems. However, while fundamental for the prediction of future activity, it remains difficult so far to determine the depth of the gas sources. Here we show how the combined measurement of CO₂ and radon-222 fluxes at the surface constrains the depth of degassing at two hydrothermal systems in geodynamically active contexts: Furnas Lake Fumarolic Field (FLFF, Azores, Portugal) with mantellic and volcano-magmatic CO₂, and Syabru-Bensi Hydrothermal System (SBHS, Central Nepal) with metamorphic CO₂. At both sites, radon fluxes reach exceptionally high values ($> 10 \text{ Bq m}^{-2} \text{ s}^{-1}$) systematically associated with large CO₂ fluxes ($> 10 \text{ kg m}^{-2} \text{ day}^{-1}$). The significant radon–CO₂ fluxes correlation is well reproduced by an advective–diffusive model of radon transport, constrained by a thorough characterisation of radon sources. Estimates of degassing depth, $2580 \pm 180 \text{ m}$ at FLFF and $380 \pm 20 \text{ m}$ at SBHS, are compatible with known structures of both systems. Our approach demonstrates that radon–CO₂ coupling is a powerful tool to ascertain gas sources and monitor active sites. The exceptionally high radon discharge from FLFF during quiescence ($\approx 9 \text{ GBq day}^{-1}$) suggests significant radon output from volcanoes worldwide, potentially affecting atmosphere ionisation and climate.

Since decades, the tenuous relation between Earth's deformation, earthquakes, and carbon dioxide (CO₂) release has fostered broad interest in the geoscience's community^{1,2}. Significant CO₂ emissions have been commonly reported at the plate boundaries in a variety of seismotectonic regimes: extension (rifting)³, reverse fault⁴, strike-slip fault⁵, subduction⁶, triple junction⁷, and collision⁸. To appreciate the potential coupling between geodynamics and CO₂, we need to better understand CO₂ sources and transport mechanisms. Besides biological sources, the released geogenic CO₂ generally has either a volcano-magmatic, a mantellic or a metamorphic source, or a mixing of these. The number of available data-sets on CO₂ release has increased⁹, and it appears timely to evaluate the available results against each other. Comparing CO₂-emitting sites in different active tectonic settings may help diagnose the involved gas transport mechanisms and constrain the depth of degassing (i.e., gas source depth). Both are delicate questions at all sites, but appear fundamental for long-term monitoring, prediction, and health risk assessment of the population. For this purpose, coupling CO₂ measurement with that of a trace gas, such as helium, mercury or radon, may reveal of utmost value.

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Radon-222, a noble, alpha-emitter radioactive gas with half-life of 3.8 days, is produced in the upper crust by alpha-decay of the solid radium-226. Tagged as a tracer of fluid transport at several active faults and hydrothermal systems worldwide, it was found sensitive to deformation and earthquakes, both in the field^{10,11} and laboratory^{12,13}, although many claimed precursory signals remain controversial¹⁴. To migrate from its source to the surface before decaying, radon needs to be transported by a carrier fluid having a sufficient velocity, such as CO₂, at high-permeability geosystems. Indeed, the use of radon to constrain gas transport and source has been pioneered at a few CO₂-emitting sites. For example, at Santorini volcano (Greece), carbon isotopic composition was compared with radon concentration, but without accounting for heterogeneity of the layers crossed by the fluid mixture¹⁵. At Mt. Etna volcano (Italy), radon transport was modelled through geological layers, but without accounting for production terms within the layers¹⁶. At Pantelleria Island (Italy) and Nisyros volcano (Greece), ²²²Rn/²²⁰Rn concentration ratios and few radon source term values in soils and rocks were used, but without radon transport modelling^{17,18}. To infer constraints on gas source and transport mechanisms by coupling CO₂ and radon measurements, a radon transport model, together with a thorough characterisation of radon sources, is necessary to interpret the data.

Here, we apply a combined approach of coupling the measurements of CO₂ flux and radon flux from the ground surface to constrain the depth of degassing at two geodynamically active sites with different CO₂ sources: Furnas Lake Fumarolic Field (FLFF), São Miguel Island, Azores archipelago, Portugal with mantlelic and volcano-magmatic CO₂ sources, and Syabru-Bensi Hydrothermal System (SBHS) in the Himalayas of Central Nepal with a metamorphic CO₂ source. We show that the combination of radon flux, CO₂ flux, radon sources, and CO₂ sources, interpreted using a common simplified multiphase radon transport model, constrains the depth of CO₂ degassing at both volcanic (FLFF) and active fault sites (SBHS), and allows quantifying deep-originated gas fluxes to the atmosphere.

The Azores archipelago, formed by nine volcanic islands along a WNW-ESE trend, is located in the North Atlantic Ocean at the triple junction of the American, Eurasian, and Nubian tectonic plates. The largest island, São Miguel, comprises three dormant trachytic polygenetic volcanoes (from west to east: Sete Cidades, Fogo, and Furnas)¹⁹. Furnas volcano (Fig. 1a), with an age of approximately 100,000 years B.P., is formed by two nested calderas controlled by a NW-SE and NE-SW trending fault system²⁰. The western part of the youngest caldera is filled by a shallow (≤ 15 m depth) lake. Several eruptive styles occurred at Furnas volcano, from effusive to caldera-forming explosive events, with a total of 10 intra-caldera eruptions in the last 5000 years²¹. Since the fifteenth century, two major eruptions have occurred: one event in 1439–1443 and one deadly subplinian event in 1630. The erupted products were mainly pumices, ashes and lapilli, surges and trachytic lava domes²⁰.

The current volcanic activity takes the form of secondary manifestations, such as boiling fumaroles, thermal and cold CO₂-rich springs, and CO₂ diffuse degassing structures (DDS). One of the main degassing areas (Fig. 1a) is Furnas Lake Fumarolic Field (FLFF) located north of Furnas Lake²². Gas released is dominated by water steam and CO₂, with detectable traces of H₂S, H₂, N₂, CH₄, Ar, He, and CO₂²². According to the recently published conceptual model, the released gases come from the final cooling of a trachytic reservoir located around 3–4 km depth²³, percolate through the fracture network, dissolve in shallow meteoric aquifers with vapour/liquid equilibrium temperature around 270 °C in Furnas Lake area, degas and then percolate to the surface^{22,24}. This is consistent with various geophysical soundings performed at FLFF, which showed P-wave low velocity zone at 6 km depth²⁵, low density magma bodies at 4–5 km depth²⁶, and conductive zones at 500 m and 100 m depth²⁷ dipping below the lake²⁸. Chemistry and isotopic composition of thermal waters and fumaroles at FLFF^{7,22,29,30} suggest mantlelic and volcano-magmatic CO₂ ($\delta^{13}\text{C} = -4.5 \pm 0.2$ ‰; $n = 3$) and a mixing of mantlelic and crustal components for helium ($R/R_a = 5.25 \pm 0.02$; $n = 3$).

The Azores climate is oceanic temperate, with a mean annual temperature of 17 °C (minimum of 14 °C in January and maximum of 25 °C in August). The mean annual precipitation of 1930 mm year⁻¹ is marked by a strong seasonality between a rainy season from October to March (75% of annual precipitation) and a dry season in summer³¹. At FLFF, CO₂ is released by fumaroles, DDS, and thermal springs. Several CO₂ flux studies^{7,32} have been carried out at the four main CO₂ emitting areas of Furnas volcano (FLFF, Furnas Village, Ribeira dos Tambores, and Ribeira Quente Village). Recent total estimated gaseous CO₂ discharge from FLFF³³ amounts to 35 t day⁻¹, among which DDS represent^{33,34} 6.0 ± 0.2 t day⁻¹ and 19.8 t day⁻¹, respectively over 4000 m² and 20,000 m² surface area. In the FLFF surroundings, a permanent CO₂ flux station is operating since 2004^{32,35}. Potential radon sources may be associated with ²²⁶Ra excess in basaltic lavas³⁶, ²²⁶Ra secondary mineralisation within ashes, pumices, lapilli, and altered soil surface, and ²²⁶Ra dissolved in thermal waters. Large radon concentrations are reported in the ground (max: 390,000 Bq m⁻³) and in habitations (max: 13,300 Bq m⁻³) at Furnas volcano^{37,38}, which, together with CO₂, pose substantial health hazard to the population^{7,39}.

The second site considered in this paper is located in the Nepal Himalayas. The Himalayan orogen results from the India–Eurasia collision which started 55 Ma ago. Half of the continent–continent shortening, 2 cm year⁻¹, is accommodated by the large décollement called the Main Himalayan Thrust (MHT)⁴⁰. In Nepal, the last large earthquake, the 2015 M_w 7.9 Gorkha earthquake, claiming >9000 lives, ruptured the MHT over 150 km⁴¹. About 4–5 events of local magnitude $M_L > 5$ occur per year, concentrated at 10–25 km depth, at the foot of the Himalayan topographic rise⁴². This area, the Main Central Thrust (MCT) zone⁴³, is a 2- to 10-km-width shear zone that separates high-grade meta-crystalline rocks from the Greater Himalaya to the north from low-grade meta-sedimentary rocks from the Lesser Himalaya to the south.

Along the whole Himalayas, the MCT zone comprises numerous hydrothermal systems characterised by thermal and cold CO₂-rich springs^{44,45}, 'tectonic' fumaroles, and CO₂ DDS⁴⁶, with significant CO₂ emissions concentrated along a 110-km-long segment in Central Nepal⁴⁷. The upper Trisuli valley (Fig. 1b), located 60 km north to Kathmandu, encompasses several CO₂-emitting hydrothermal sites in the vicinity of the MCT zone and shows the largest Himalayan CO₂ release reported so far ($> 15 \pm 3$ t day⁻¹)^{47,48}. This valley also reported significant post-seismic hydrothermal changes following the Gorkha earthquake⁸. One of these sites, Syabru-Bensi Hydrothermal

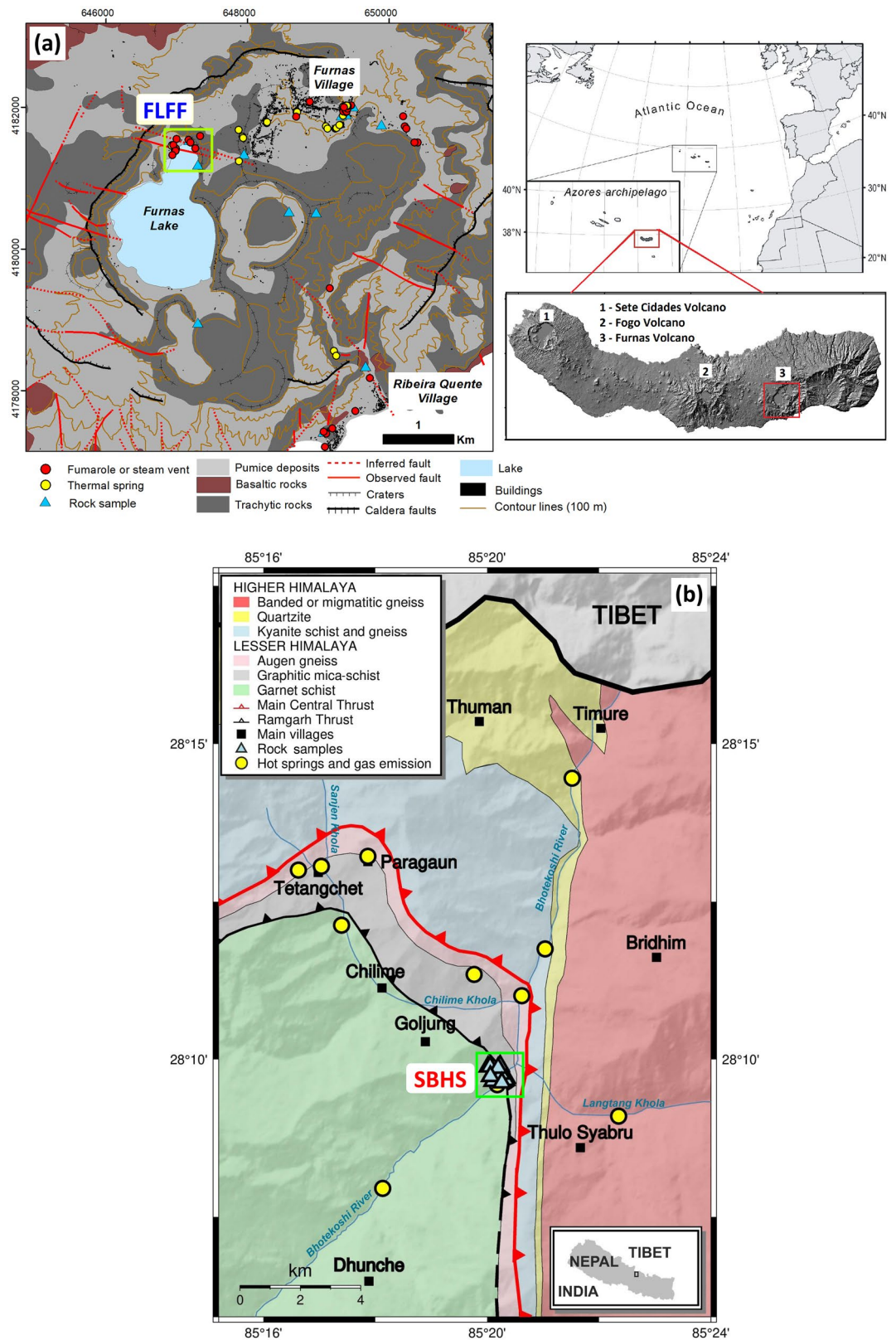


Figure 1. (a) Geological map showing the location of the Furnas Lake Fumarolic Field (FLFF) in mantellic and volcano-magmatic context (Azores, Portugal). The insets show the location of the Azores and of the Furnas Volcano on the São Miguel Island. (b) Geological map showing the location of the Syabru-Bensi Hydrothermal System (SBHS) in metamorphic and active tectonic context (Central Nepal). The inset shows the location in the Nepal map.

System (SBHS) (Fig. 1b), is located 3.5 km south to MCT within the Lesser Himalayan rocks comprising mainly mica-schist, quartzite, marble, graphitic schist, and augen gneiss^{49–51}. Gas released is dominated by CO₂, with water steam and large H₂S content (340 ppm). According to the most recent conceptual model⁸, CO₂-rich gas comes from a deep source at > 5 km depth, percolates through fracture networks in the MCT zone, forms gas reservoirs and mixes with meteoric water, degasses at a depth of 100 m or more at 70 °C ≤ T ≤ 120 °C (vapour/liquid equilibrium temperature), and then crosses near-surface aquifers before reaching the surface. This appears consistent with geophysical soundings carried out at SBHS and in the valley which showed a highly conductive and low P-wave velocity zone at ≈ 10 km depth⁵², conductive zones at 10–30 m depth and shallow altered fractured conduits for gas below the SBHS⁵³, and self-potential anomalies at the surface^{53,54}. Chemistry and isotopic composition of thermal waters and DDS at SBHS^{8,44,45,54} suggest CO₂ production by metamorphic decarbonation at > 5 km depth, a dominating crustal carbon component of the gaseous CO₂ ($\delta^{13}\text{C} = -0.75 \pm 0.01\text{‰}$; $n = 27$), and radiogenic helium ($R/R_a \leq 0.05$; $n = 2$).

Circumvented respectively to the west, north, and east by Ganesh, Tibet, and Langtang ranges, SBHS benefits from a rain shadow effect of Gosainkunda range to the southeast. The mean annual precipitation varies from 1100 to 1800 mm year⁻¹. Monsoon occurs from June to September (80% of annual precipitation) and dry season from December to February⁵⁵. The mean annual air temperature is 19 °C, with minimum of 0 °C in January and maximum of 28 °C in June. At SBHS, CO₂ is released by fumaroles, DDS, and thermal springs. Several CO₂ flux studies^{8,46,47,54,56} have been carried out at the five main CO₂-emitting areas of SBHS. Total estimated gaseous CO₂ discharge from SBHS^{8,47,54} amounts to 8 t day⁻¹, among which 3.8 ± 0.4 t day⁻¹ corresponds to DDS of gas zones 1 and 2 over a surface area of 11,000 m². At this site, several radon measurement campaigns reported large radon fluxes at the surface (max: 38,500 × 10⁻³ Bq m⁻² s⁻¹) and radon concentrations in the ground (max: 57,700 Bq m⁻³)^{8,46,54,57}. Radon sources have also been investigated in the thermal and cold waters, the soil, and the surrounding rock layers^{50,54,58}. These high CO₂ and radon releases also pose a health hazard to the local population and animals^{54,57}.

Results

Radon and CO₂ fluxes at Furnas Fumarolic Field. Radon and CO₂ fluxes were measured in June 2016 at the surface using the accumulation chamber method (see “Methods” section; Supp. Fig. S1) around the main boiling fumaroles and near the lakeshore. A total of 169 radon and 371 CO₂ flux measurements were performed at 136 and 335 locations (Table 1), respectively, over a surface area of 48,000 m². Fluxes were measured along several profiles every 2 to 5 m (Supp. Fig. S2).

Mean radon flux values range over five orders of magnitude from 1.3 to 40,000 × 10⁻³ Bq m⁻² s⁻¹ (Fig. 2a). The largest radon fluxes (> 10,000 × 10⁻³ Bq m⁻² s⁻¹) are found in the fumarolic area. Such huge values are commonly reported on uranium mill tailings (see review⁵⁴). The overall arithmetic (geometric) mean for radon fluxes amounts to 4300 ± 600 (560 ± 10) × 10⁻³ Bq m⁻² s⁻¹, which is more than three times larger than the few reported mean radon flux values obtained at volcanic sites worldwide (see review⁵⁴). The distribution of radon flux values is bimodal (Fig. 2a), with populations A (55%) and B (45%) separated by a threshold value of 870 × 10⁻³ Bq m⁻² s⁻¹ using normal probability partitioning⁵⁹ (Table 1); the median (720 ± 20 × 10⁻³ Bq m⁻² s⁻¹) is larger than the mean. Mean CO₂ flux values range over five orders of magnitude from 8.1 to 27,000 g m⁻² day⁻¹ (Fig. 2c), with the largest values (> 1000 g m⁻² day⁻¹) also found in the fumarolic area. The overall arithmetic (geometric) mean for CO₂ fluxes amounts to 1200 ± 200 (139 ± 1) g m⁻² day⁻¹, compatible with past measurement campaigns^{7,33,34}. The distribution of CO₂ flux values is bimodal, with populations A (81%) and B (19%) separated by a threshold value of 720 g m⁻² day⁻¹ (Table 1); the median (83.3 ± 0.2 g m⁻² day⁻¹) is smaller than the mean.

Based on the whole flux data-set and using sequential Gaussian simulations (see “Methods” section), maps of radon and CO₂ fluxes are obtained at FLFF (Fig. 3). The largest radon fluxes are concentrated in the fumarolic area, near the parking, and on the eastern part, while the lowest are found in the western lakeshore and between the fumarolic area and the parking (Fig. 3a). Significant spatial variations of radon flux are found within the fumarolic area over an area of 80 × 100 m² (inset of Fig. 3a), with difference of several orders of magnitude within few metres only. On the eastern part, large radon fluxes appear isolated. The largest CO₂ fluxes are also concentrated in the fumarolic area, near the parking and on the eastern part, while the lowest are found in the western lakeshore and between the fumarolic area and the parking (Fig. 3b). Significant spatial variations of CO₂ flux, larger than for radon fluxes, are found within the fumarolic area (inset of Fig. 3b), with difference of several orders of magnitude within few metres only. On the eastern and western parts and near the parking, large CO₂ fluxes reflect areas with a higher mean CO₂ flux. The surface area of large CO₂ flux is larger than that of large radon flux.

Based on our data-set, past CO₂ flux campaigns⁷, and the literature, mean background fluxes are estimated as 22 × 10⁻³ Bq m⁻² s⁻¹ for radon and 25 g m⁻² day⁻¹ for CO₂. The surface areas of radon and CO₂ fluxes above background yield 43,700 and 45,200 m², respectively (about 97% of the investigated surface area). Exceptionally large radon (> 1000 × 10⁻³ Bq m⁻² s⁻¹) and CO₂ fluxes (> 1000 g m⁻² day⁻¹) occupy a surface area of 23,900 m² (54%) and 3600 m² (8%), respectively. The total estimated radon discharge amounts to 9300 ± 1600 MBq day⁻¹ (108,000 ± 19,000 Bq s⁻¹). To date, this huge estimate is the first obtained at a volcanic hydrothermal system. The total estimated CO₂ discharge amounts to 19.1 ± 5.3 t day⁻¹ (5.0 ± 1.4 mol s⁻¹). This value, consistent with reported CO₂ discharges obtained during past campaigns^{7,34}, is similar to other volcanic sites worldwide⁴⁷.

Radon and CO₂ fluxes at Syabru–Bensi hydrothermal system. Radon and CO₂ fluxes were regularly measured from 2009 to 2020 at the surface using the accumulation chamber method (see “Methods” section; Supp. Fig. S1) around the tectonic fumaroles and non-vegetated areas, 20 m above the main hot springs (gas zones 1–2). A total of 418 radon and 777 CO₂ flux measurements were performed at 250 and 399 locations

Site	FLFF	SBHS
Radon-222 flux (10^{-3} Bq m$^{-2}$ s$^{-1}$)		
Investigated surface area (m 2)	44,350	948
Number of measurements	169	418
Number of measured points	136	250
Min–max range	1.34–39,558	1.01–20,476
Arithmetic mean	4302 \pm 636	1142 \pm 139
Geometric mean	562 \pm 13	233.6 \pm 3.4
Median (at 90% CL)	722 \pm 18	225.8 \pm 9.9
Min–max range (at 90% CL)	6.6–22,780	12.1–4999
Normal probability partitioning		
Population A: min–max (fraction)	1.3–870 (54.7%)	1.0–130 (43.0%)
Population B: min–max (fraction)	870–40,000 (45.3%)	130–20,000 (57.0%)
Estimated total discharge (MBq d $^{-1}$)	9300 \pm 1600	71 \pm 11
Estimated total discharge (Bq s $^{-1}$)	108,000 \pm 19,000	820 \pm 130
Estimated surface of flux > 1000 $\times 10^{-3}$ Bq m $^{-2}$ s $^{-1}$ (m 2)	23,864	172
Estimated surface of flux > 500 $\times 10^{-3}$ Bq m $^{-2}$ s $^{-1}$ (m 2)	32,532	411
Estimated surface of flux > 300 $\times 10^{-3}$ Bq m $^{-2}$ s $^{-1}$ (m 2)	36,405	589
Estimated surface of flux > 100 $\times 10^{-3}$ Bq m $^{-2}$ s $^{-1}$ (m 2)	41,118	856
Estimated surface of flux > 22 $\times 10^{-3}$ Bq m $^{-2}$ s $^{-1}$ (m 2)	43,654	937
CO$_2$ flux (g m$^{-2}$ day$^{-1}$)		
Investigated surface area (m 2)	47,745	3078
Number of measurements	371	777
Number of measured points	335	399
Min–max range	8.1–26,698	2.76–155,178
Arithmetic mean	1227 \pm 184	6463 \pm 1007
Geometric mean	139.4 \pm 0.9	284.0 \pm 2.7
Median (at 90% CL)	83.3 \pm 0.2	173.6 \pm 1.4
Min–max range (at 90% CL)	15.8–7126	10.1–46,773
Normal probability partitioning		
Population A: min–max (fraction)	8.1–720 (80.8%)	2.8–260 (54.8%)
Population B: min–max (fraction)	720–27,000 (19.2%)	260–7800 (33.9%)
Population C: min–max (fraction)		7800–160,000 (11.3%)
Estimated total discharge (t day $^{-1}$)	19.1 \pm 5.3	6.3 \pm 1.6
Estimated total discharge (mol s $^{-1}$)	5.0 \pm 1.4	1.65 \pm 0.41
Estimated surface of flux > 1000 g m $^{-2}$ day $^{-1}$ (m 2)	3638	890
Estimated surface of flux > 500 g m $^{-2}$ day $^{-1}$ (m 2)	4070	1239
Estimated surface of flux > 300 g m $^{-2}$ day $^{-1}$ (m 2)	4164	1536
Estimated surface of flux > 100 g m $^{-2}$ day $^{-1}$ (m 2)	24,428	2178
Estimated surface of flux > 25 g m $^{-2}$ day $^{-1}$ (m 2)	45,192	2810
Estimated surface of flux > 10 g m $^{-2}$ day $^{-1}$ (m 2)	47,743	3045
Radon–CO$_2$ fluxes correlation		
Number of averaged values	136	157
Correlation coefficient	0.83 \pm 0.02	0.80 \pm 0.03

Table 1. Overview of radon and CO $_2$ flux data-set separately at Furnas Lake Fumarolic Field (FLFF) and Syabru-Bensi Hydrothermal System (SBHS). Estimated surface discharge of radon and CO $_2$ are also given for both sites. All experimental uncertainties are given at one standard deviation (1- σ , 68% confidence level).

(Table 1), respectively, over a surface area of 3000 m 2 . Fluxes were measured along several profiles every 1 to 2 m (Supp. Fig. S3).

Mean radon flux values range over five orders of magnitude from 1.0 to 20,000 $\times 10^{-3}$ Bq m $^{-2}$ s $^{-1}$ (Fig. 2b). The largest radon fluxes (> 10,000 $\times 10^{-3}$ Bq m $^{-2}$ s $^{-1}$) are concentrated near the tectonic fumaroles. The overall arithmetic (geometric) mean for radon fluxes amounts to 1100 \pm 100 (234 \pm 3) $\times 10^{-3}$ Bq m $^{-2}$ s $^{-1}$. The distribution of radon flux values is bimodal (Fig. 2b), with populations A (43%) and B (57%) separated by a threshold value of 130 $\times 10^{-3}$ Bq m $^{-2}$ s $^{-1}$ (Table 1); the median (230 \pm 10 $\times 10^{-3}$ Bq m $^{-2}$ s $^{-1}$) is similar to the mean. Mean CO $_2$ flux values range over six orders of magnitude from 2.8 to 160,000 g m $^{-2}$ day $^{-1}$ (Fig. 2d). The largest CO $_2$ fluxes (> 1000 g m $^{-2}$ day $^{-1}$) are also measured in the vicinity of the fumaroles. The overall arithmetic (geometric) mean for CO $_2$ fluxes amounts to 6500 \pm 1000 (284 \pm 3) g m $^{-2}$ day $^{-1}$. The distribution of CO $_2$ flux values bears

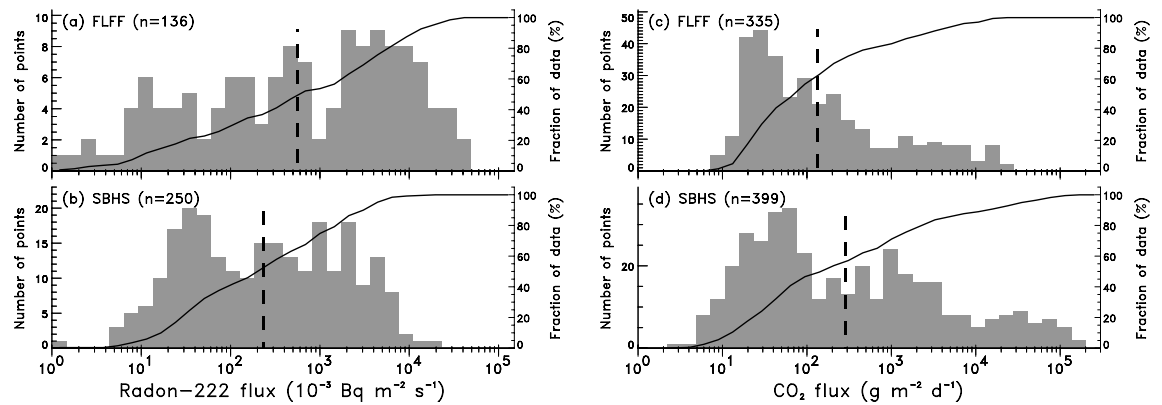


Figure 2. Distributions of radon and CO₂ fluxes measured at both sites (in logarithmic scale): on the left, radon flux at (a) FLFF and (b) SBHS; on the right, CO₂ flux at (c) FLFF and (d) SBHS. Geometric mean of each distribution is represented as a vertical dashed black line and the cumulated distribution as a solid black curve (scale on the right-hand side).

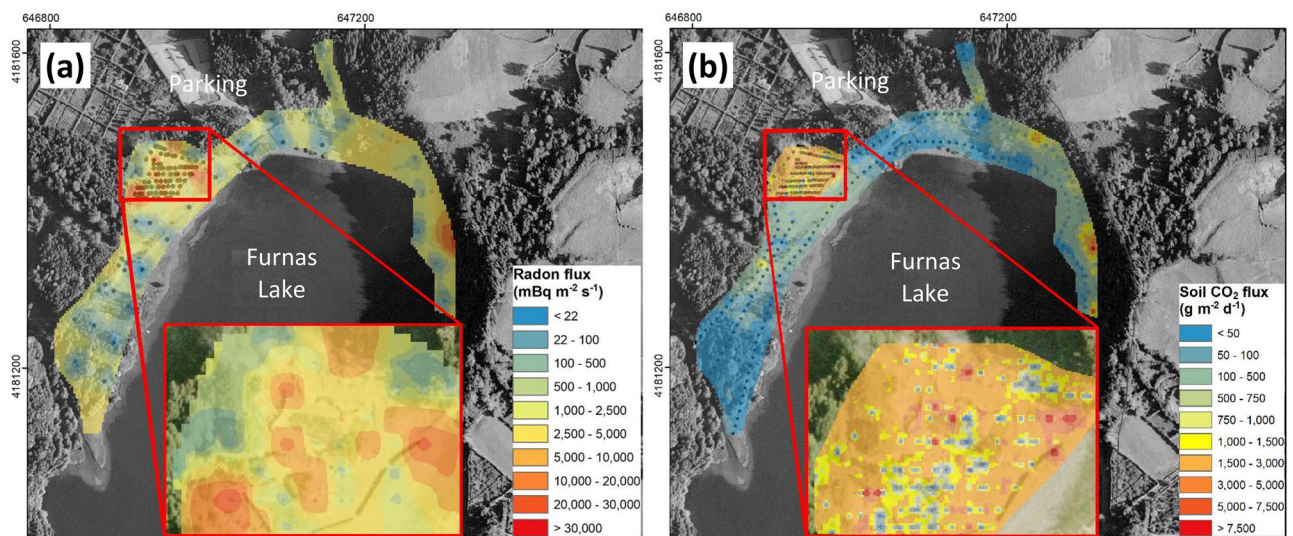


Figure 3. Interpolated (a) radon and (b) CO₂ flux maps of FLFF. For each flux map, the colour scale is shown on the bottom right. In (a,b), the inset is an enlargement of the fumarolic area. The map is projected following the UTM coordinate system. The map was built using sGs method with a cell size of 3 m² (see “Methods” section).

three modes (Fig. 2d), with populations A (55%), B (34%) and C (11%) separated by threshold values of 260 and 7800 g m⁻² day⁻¹, respectively (Table 1); the median (174 ± 1 g m⁻² day⁻¹) is smaller than the mean.

Similarly (see “Methods” section), maps of radon and CO₂ fluxes are obtained at SBHS (Fig. 4). The largest radon fluxes are concentrated in the vicinity of small recesses at the base of terrace scarps, in non-vegetated areas, and at the foot of the terraces arranged for crops, while the lowest are found in the southern and northern parts of the alluvial and debris fall terrace (Fig. 4a). Here also, metre-scaled variations of radon flux of several orders of magnitude are noticed around recesses and non-vegetated areas over a surface area of 30×45 m² (Fig. 4a). In the western area, large radon fluxes are found in the vicinity of inhabited housings. The largest CO₂ fluxes are also concentrated inside and near the recesses, in non-vegetated areas, and at the foot of terraces, while the lowest are found in the southern and northern parts (Fig. 4b). Similarly, significant spatial variations of CO₂ flux are found near the recesses and non-vegetated areas (Fig. 4b), with difference of several orders of magnitudes within few metres only. Here also, the surface area of large CO₂ flux is larger than that of large radon flux.

Based on the overall data-set, mean background fluxes are estimated as 22×10^{-3} Bq m⁻² s⁻¹ for radon and 10 g m⁻² day⁻¹ for CO₂. The surface areas of radon and CO₂ fluxes above background yield 940 m² and 3000 m², respectively (about 99% of the investigated surface area). Exceptionally large radon ($> 1000 \times 10^{-3}$ Bq m⁻² s⁻¹) and CO₂ fluxes (> 1000 g m⁻² day⁻¹) occupy a surface area of 170 m² (18%) and 890 m² (29%), respectively. The total estimated radon discharge amounts to 71 ± 11 MBq day⁻¹ (820 ± 130 Bq s⁻¹). The total estimated CO₂ discharge amounts to 6.3 ± 1.6 t day⁻¹ (1.65 ± 0.41 mol s⁻¹), similar to previously published values at this site^{8,54} and to other mofette sites worldwide⁴⁷.

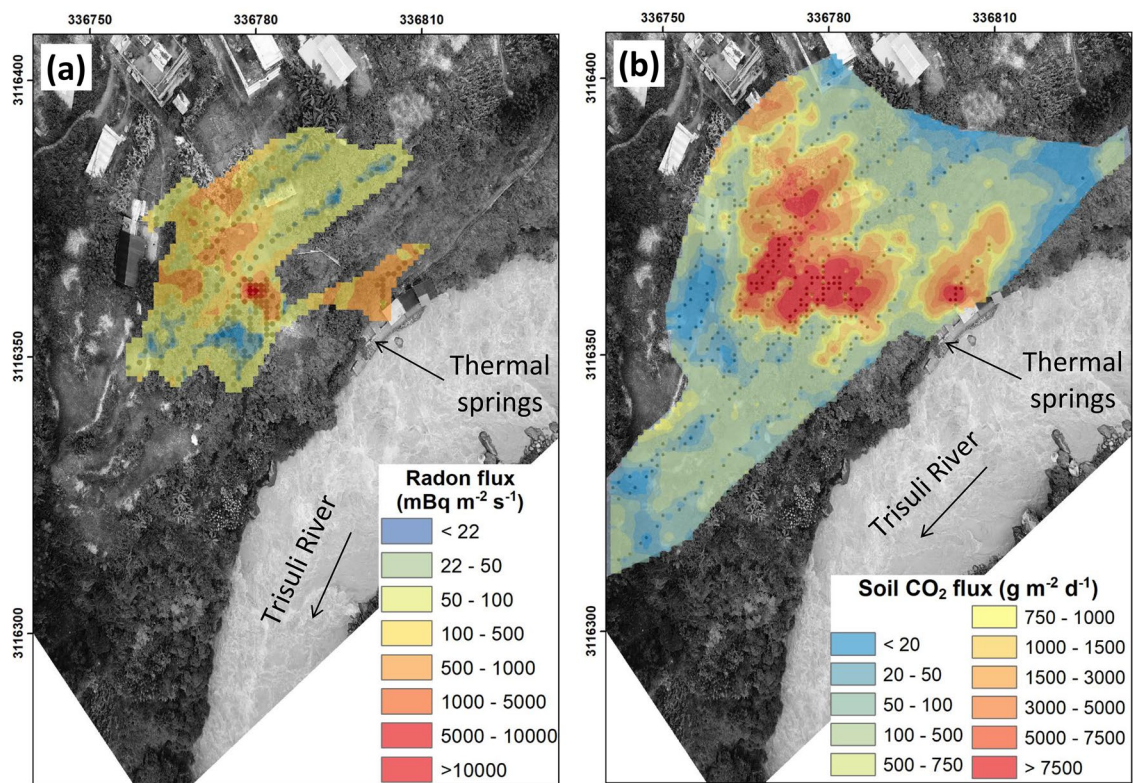


Figure 4. Interpolated (a) radon and (b) CO₂ flux maps of SBHS. For each flux map, the colour scale is shown on the bottom right. The map is projected following the UTM coordinate system. The map was built using sGs method with a cell size of 1 m² (see “Methods” section).

Radon sources. At FLFF and SBHS, numerous potential sources of radon were investigated (Table 2). At FLFF, 45 soil and 18 rock samples were analysed for their effective radium-226 concentration (EC_{Ra} , or radon source term; see “Methods” section). Soil samples consist mostly of altered surface rich in kaolinite and marcasite²⁹ in the fumarolic area ($n=30$), and of pumice volcanic material of the latest explosive eruption³⁵ outside ($n=15$). Soil EC_{Ra} values range from 2.3 to 21 Bq kg⁻¹, with an arithmetic (geometric) mean of 8.6 ± 0.7 (7.4 ± 0.1) Bq kg⁻¹ (Supp. Fig. S4). This value is consistent with the mean EC_{Ra} value for soils^{60,61} (≈ 7 Bq kg⁻¹; $n=2070$). EC_{Ra} values for volcanic rocks from samples collected inside and around the Furnas caldera, i.e. altered rock and pumice (lapilli and ash), range from 0.3 to 34 Bq kg⁻¹, with arithmetic (geometric) mean of 6.2 ± 2.0 (3.2 ± 0.1) Bq kg⁻¹ (Supp. Fig. S4), significantly larger than the mean EC_{Ra} value for volcanic rocks⁶⁰ (1.7 ± 0.2 Bq kg⁻¹; $n=349$; Supp. Fig. S5a), suggesting a crustal signature.

Radon concentration in ground gas, measured at 86 locations (see “Methods” section), ranges from 179 to 377,000 Bq m⁻³, with a mean of $33,100 \pm 7200$ Bq m⁻³, consistent with reported values^{37,38} and mean radon concentration in water bubbling ($74,000 \pm 18,000$ Bq m⁻³). By contrast, radon concentration in the boiling fumarole is low (810 ± 31 Bq m⁻³). Mean radium-226 and radon concentrations in thermal water (see “Methods” section) are low (means of $63 \pm 32 \times 10^{-3}$ Bq L⁻¹ and 4.1 ± 0.2 Bq L⁻¹, respectively), in the lower range of values for hydrothermal waters⁶².

At SBHS, 68 soil and 19 rock samples give arithmetic mean (min–max) EC_{Ra} values of 9.2 ± 0.7 (1.0–43) and 2.5 ± 0.5 (0.13–7.5) Bq kg⁻¹, respectively (Supp. Fig. S4). Soil consists of debris fall deposit of mica-schist mixed with alluvial soil in the vegetated area, and of hydrothermal soil (“reduktosol” or “mofettic” qualification⁶³) richer in organic matter, clay, and secondary iron oxides⁵⁴ in the non-vegetated area. The metamorphic rocks are mainly garnet-rich mica-schist, marble, and quartzite. Their EC_{Ra} values are relatively similar to the mean EC_{Ra} value for metamorphic rocks (5.1 ± 0.4 Bq kg⁻¹; $n=1256$; Supp. Fig. S5b). Mean soil EC_{Ra} is similar to that obtained at FLFF, but mean rock EC_{Ra} is smaller.

Radon concentration in ground gas, measured at six locations, yields mean of $43,000 \pm 1800$ Bq m⁻³. Bubbling waters, only reported on the opposite river bank, give smaller radon concentration in water bubbles (mean: $11,300 \pm 1600$ Bq m⁻³). Similarly, mean radium-226 and radon concentrations in thermal waters are not exceptional (means of $125 \pm 66 \times 10^{-3}$ Bq L⁻¹ and 15 ± 9 Bq L⁻¹, respectively).

CO₂ sources. At FLFF, ground gas CO₂ concentration, measured at 89 locations (see “Methods” section), ranges from 0.3 to ≈ 100 vol% (mean: $57 \pm 4\%$), consistent with reported values^{7,39}. Total dissolved inorganic carbon (DIC) concentration in thermal waters gives $9.0 \pm 0.9 \times 10^{-3}$ mol L⁻¹. The carbon isotopic composition of CO₂, with mean $\delta^{13}C$ of $-4.2 \pm 0.3\%$ for gaseous CO₂ and -4% for DIC, confirms the mantellic and volcano-magmatic CO₂ sources (about -4%) at FLFF^{7,22}.

Site		FLFF	SBHS
Radon-222 sources			
Soil EC_{Ra} (Bq kg ⁻¹)	Number of soil samples	45	68
	Min–max range	2.30–20.9	1.03–42.9
	Min–max range (at 90% CL)	2.80–16.7	2.02–15.8
	Arithmetic mean	8.56 ± 0.69	9.15 ± 0.72
	Geometric mean	7.44 ± 0.07	7.68 ± 0.06
Rock EC_{Ra} (Bq kg ⁻¹)	Number of rock samples	18	19
	Min–max range	0.27–33.6	0.13–7.46
	Min–max range (at 90% CL)	0.27–33.6	0.13–7.46
	Arithmetic mean	6.2 ± 2.0	2.47 ± 0.49
	Geometric mean	3.24 ± 0.05	1.55 ± 0.08
Ground gas radon-222 concentration (Bq m ⁻³)	Number of sample points	86	6
	Min–max range	179–376,500	39,000–48,700
	Min–max range (at 90% CL)	1530–152,000	39,000–48,700
	Mean ± 1σ	33,106 ± 7246	43,017 ± 1795
Water bubbling radon-222 concentration (Bq m ⁻³)	Number of water bubbling samples	3	2 (other springs)
	Range min–max	43,150–105,361	9137–13,559
	Mean ± 1σ	74,300 ± 18,000	11,348 ± 1563
Water radium-226 concentration (10 ⁻³ Bq L ⁻¹)	Number of springs	2	2
	Range min–max	19–107	32–218
	Mean ± 1σ	63 ± 32	125 ± 66
Water radon-222 concentration (Bq L ⁻¹)	Number of springs	2	2
	Range min–max	3.9–4.3	2.6–27.3
	Mean ± 1σ	4.10 ± 0.18	15.1 ± 8.8
CO₂ sources			
Ground gas CO ₂ concentration (%)	Number of sample points	86	4
	Min–max range	0.3–99.9	90.0–98.1
	Min–max range (at 90% CL)	0.7–99.9	90.0–98.1
	Mean ± 1σ	56.8 ± 4.3	95.4 ± 1.9
δ ¹³ C of gaseous CO ₂ (‰)	Number of sample points	6 ^a	4
	Range min/max	–4.80/3.32	–0.768/–0.713
	Mean ± 1σ	–4.35 ± 0.22	–0.746 ± 0.011
Dissolved inorganic carbon content (10 ⁻³ mol L ⁻¹)	Number of water springs	1 ^b	2
	Range min–max		29.58–32.26
	Mean ± 1σ	9.0 ± 0.9	30.92 ± 0.95
δ ¹³ C of dissolved CO ₂ (‰)	Number of water springs	1 ^b	2
	Range min/max		–0.063/0.968
	Mean ± 1σ	ca. –4	0.45 ± 0.36
Gas temperature			
Surface temperature (°C)	Number of sample points	335	31 ^c
	Min–max range	17.3–98.3	8.2–24.7
	Min–max range (at 90% CL)	18.2–52.5	8.2–24.7
	Mean ± 1σ	29.55 ± 0.75	16.31 ± 0.77
Ground temperature (°C)	Number of sample points	85	8
	Min–max range	17.1–99.4	21.6–28.8
	Min–max range (at 90% CL)	18.1–97.9	21.6–28.8
	Mean ± 1σ	45.8 ± 3.0	25.78 ± 0.87

Table 2. Overview of radon and CO₂ sources data-set separately at FFLF and SBHS. Surface and ground gas temperature is also given for both sites. ^aInclude four original data, one value from Ref.⁷ and one value from Ref.²². ^bData from Ref.²⁴. ^cInferred from surface temperature data used to estimate surface heat fluxes from Ref.⁵⁴.

At SBHS, ground gas CO₂ concentration, measured in the fumaroles, gives a high mean value of 95 ± 2%. Thermal waters have more DIC than at FFLF, with 31 ± 1 × 10⁻³ mol L⁻¹ on average. Mean δ¹³C of –0.72 ± 0.01‰ for gaseous CO₂ and 0.5 ± 0.4‰ for DIC, confirms a metamorphic CO₂ source (from 0 to –2‰) at SBHS⁴⁶. Finally, surface and ground gas temperatures are higher at FFLF compared with SBHS (Table 2).

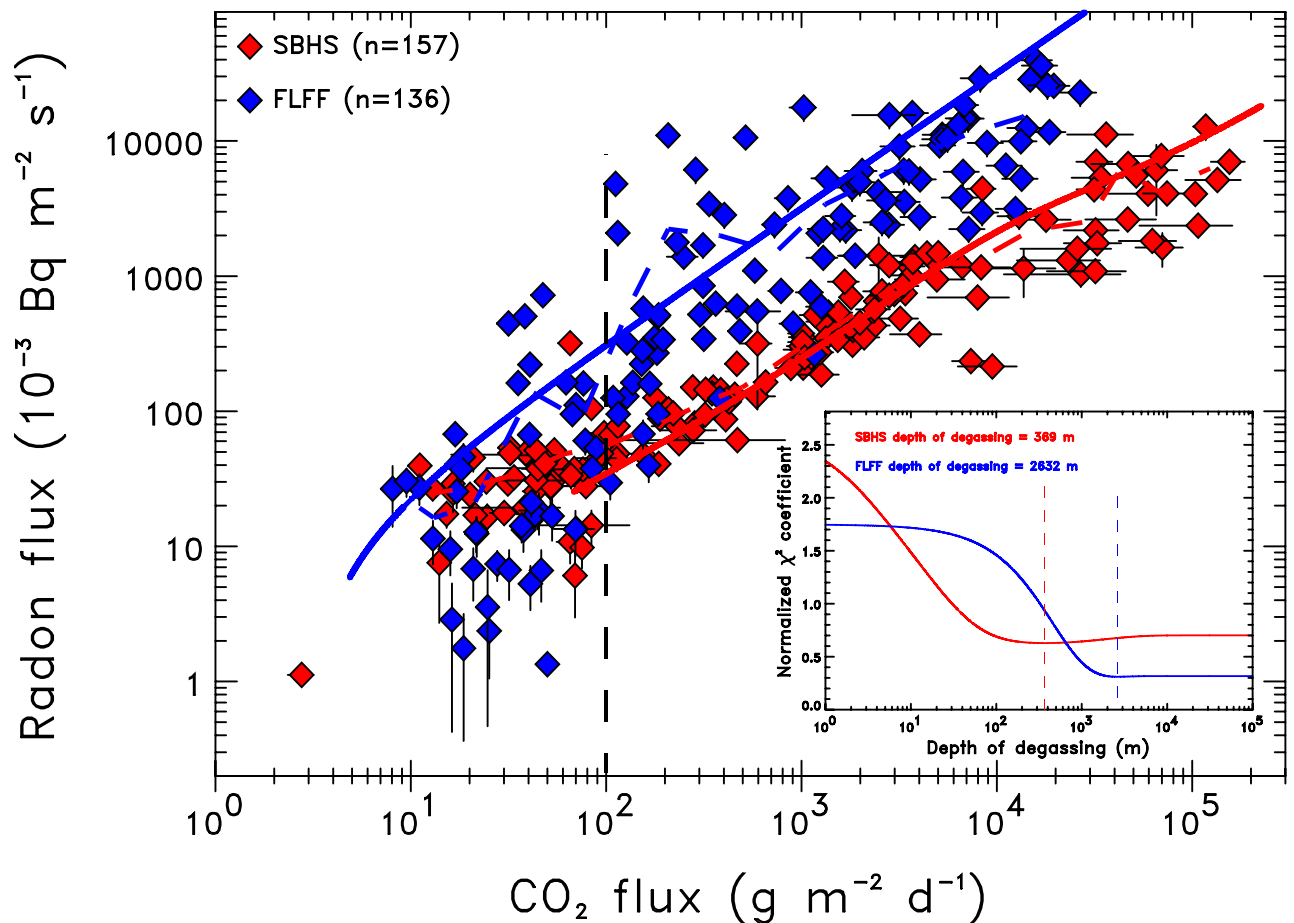


Figure 5. Radon–CO₂ fluxes correlation for FLFF (in blue) and SBHS (in red). Only fluxes measured after the 2015 Gorkha earthquake are considered at SBHS. Diamond shows the data-set, dashed curve represents the average of the data, and solid curve is the calculation of the advective–diffusive radon transport model separately for each site using the depth of degassing constrained by an unbiased approach with 500 simulations (see “Methods” section). The bottom right inset represents the normalized χ^2 coefficient as a function of the depth of the radon source, constraining the depth of CO₂ degassing separately for each site (369 m for SBHS and 2632 m for FLFF). Model parameters are summarised in Table 3.

Radon–CO₂ fluxes correlation. A general spatial agreement is found between radon flux and CO₂ flux patterns both at FLFF (Supp. Fig. S2) and SBHS (Supp. Fig. S3). Combined radon and CO₂ fluxes were measured exactly at the same point at 136 and 157 locations at FLFF and SBHS, respectively (see “Methods” section). High radon–CO₂ fluxes correlation coefficients of 0.83 ± 0.02 for FLFF and 0.80 ± 0.03 for SBHS suggest that CO₂ is the main carrier gas of radon at both sites. Radon and CO₂ fluxes are indeed strongly correlated over more than five orders of magnitude at both sites (Fig. 5), following a power-law relationship: $\Phi_{Rn} = 0.974\Phi_{CO_2}^{1.053}$ ($R^2 = 0.78$) for FLFF, and $\Phi_{Rn} = 2.441\Phi_{CO_2}^{0.681}$ ($R^2 = 0.90$) for SBHS. Despite a high dispersion of flux values for each site, systematic differences are larger than the dispersion and both sites can be discriminated. For a given CO₂ flux, larger radon flux is observed at FLFF compared with SBHS, compatible with the larger radon source term of rocks at FLFF. The larger dispersion for FLFF may be explained by the larger range of radon source terms. As radon is carried by CO₂ to the surface, radon transport mechanism is dominated by diffusion when CO₂ flux remains small ($< 100 \text{ g m}^{-2} \text{ day}^{-1}$), and by advection when CO₂ flux increases^{47,56,58}. For CO₂ fluxes $> 100 \text{ g m}^{-2} \text{ day}^{-1}$ in the advective domain (Fig. 5), radon fluxes are systematically larger at FLFF, and reach particularly high values for CO₂ flux $> 1000 \text{ g m}^{-2} \text{ day}^{-1}$, while, at SBHS, they tend to saturate. This confirms the reliability of our approach and suggests a shallower gas source at SBHS, a concept which is confirmed below by a detailed calculation.

To interpret further Fig. 5, indeed, we consider a simplified model, less complex than the reality of the two sites, but able to reproduce the essence of the radon signature of CO₂. We use an updated advective–diffusive transport model of radon carried upward by CO₂ to the surface⁵⁸ (see “Methods” section; Supp. Fig. S6). This model considers three phases for radon (gaseous, dissolved, and adsorbed) and three layers (soil, rock, and deep rock), and was modified to include temperature- and water-saturation-dependence of several radon parameters, including radon source term, an important modification for hydrothermal and volcanic sites. Since most parameters are assessed from field and laboratory data-sets, the model is able to represent both sites in term of radon transport from source(s) to surface. Deep rock layer is assumed saturated and its temperature is set to the vapour/liquid equilibrium temperature. For each layer, fixing porosity, water saturation, gas temperature, and radon source term, the depth of the deepest interface (rock – deep rock interface) can be calculated using

Site	FLFF	SBHS
Soil layer		
Thickness (m)	1	1
Porosity (%)	10	25
Water saturation (%)	80	50
Gas temperature (°C)	29.6 (29.6 ± 0.8; 500 simulations)	16.3 (16.3 ± 0.8; 500 simulations)
Water–air partition coefficient	0.20 (recalculated; 500 simulations)	0.29 (recalculated; 500 simulations)
Adsorption coefficient (m ³ kg ⁻¹)	4.0 × 10 ⁻⁵ (recalculated; 500 simulations)	5.7 × 10 ⁻⁵ (recalculated; 500 simulations)
Corrected EC _{Ra} (Bq kg ⁻¹)	6.6 (6.6 ± 0.5; 500 simulations)	8.1 (8.1 ± 0.6; 500 simulations)
Rock layer		
Constrained thickness (m)	2632 (2580 ± 180; 500 simulations)	369 (380 ± 20; 500 simulations)
Porosity (%)	5	5
Water saturation (%)	80	50
Gas temperature (°C)	45.8 (45.7 ± 3.0; 500 simulations)	25.8 (25.7 ± 0.9; 500 simulations)
Water–air partition coefficient	0.15 (recalculated; 500 simulations)	0.22 (recalculated; 500 simulations)
Adsorption coefficient (m ³ kg ⁻¹)	3.8 × 10 ⁻⁵ (recalculated; 500 simulations)	6.3 × 10 ⁻⁵ (recalculated; 500 simulations)
Corrected EC _{Ra} (Bq kg ⁻¹)	5.0 (5.0 ± 0.9; 500 simulations)	2.3 (2.3 ± 0.5; 500 simulations)
Deep rock layer		
Porosity (%)	5	5
Water saturation (%)	100	100
Gas temperature (°C)	270	120
Water–air partition coefficient	0.1	0.11
Adsorption coefficient (m ³ kg ⁻¹)	3.8 × 10 ⁻⁵	6.3 × 10 ⁻⁵
Corrected EC _{Ra} (Bq kg ⁻¹)	22	0.58

Table 3. Parameters used in the advective–diffusive model of radon transport separately for FLFF and SBHS. The thickness of the rock layer, constrained by 500 simulations, gives an estimate of the depth of degassing (see text).

published equations⁵⁸ (see “Methods” section; Table 3). We optimise this interface depth using a normalised χ^2 coefficient calculated for each data-set (i number of data with CO₂ flux > 100 g m⁻² day⁻¹), $\chi^2 = \sum_i (\Phi_{Rn,i}^{meas} - \Phi_{Rn,i}^{calc})^2 / (\sigma_{\Phi_{Rn,i}}^2 + \sigma_{\Phi_{CO_2,i}}^2)$, where, respectively, Φ_{Rn}^{meas} and Φ_{Rn}^{calc} are measured and calculated radon fluxes, and $\sigma_{\Phi_{Rn}}$ and $\sigma_{\Phi_{CO_2}}$ are one-sigma uncertainty on radon and CO₂ fluxes. This interface corresponds to the maximum depth of the radon source, where CO₂ velocity is high enough to carry radon before its decay, and therefore represents the maximum depth of radon-carrier degassed CO₂. In some cases, it corresponds to the maximum depth of CO₂ degassing. In other cases, CO₂ degassing can be deeper and this depth can be seen as a minimum depth for degassed CO₂.

Gas temperature and radon source term having larger effects on degassing depth (“Methods” section; Supp. Fig. S7), we vary these parameters of the soil and rock layers around the mean with 500 simulations. Minimum normalised χ^2 coefficients give an optimized depth of degassing for each simulation, and the median of their distribution (Supp. Fig. S8) yields 2580 ± 180 m for FLFF and 380 ± 20 m for SBHS (Table 3 and inset of Fig. 5). Injecting these constrained depths, the model reproduces well the radon–CO₂ fluxes correlation over four to five orders of magnitude (Fig. 5). Furthermore, these depths match the general overview of CO₂ transport at both sites, as described above. At FLFF, degassing depth of 2.6 km is compatible with the presence of a crystallised body of mica-rich syenite, which previous studies attributed a depth of 3–4 km²³. At SBHS, degassing depth of 380 m is consistent with a shallow CO₂ reservoir, sensitive to crustal deformation and earthquakes⁸. While the interpretation of the estimated depth of degassing, when it is large, needs to be cautious, the one-order-of-magnitude difference in degassing depth between FLFF and SBHS is evidenced without doubt. This difference could also be examined through the prism of the CO₂ source, as volcanic settings with mantellic and volcano-magmatic CO₂ might have deeper roots than collisional settings with metamorphic CO₂.

Discussion

The results shown above indicate that the most likely source of radon is from the rock. However, other radon sources have been considered: the thermal waters and the soil. At both sites, a water-degassing model⁵⁸ (Supp. Fig. S9; see “Methods” section) or a purely diffusive model⁵⁸ (Supp. Fig. S10; see “Methods” section) cannot

account for the obtained estimated radon and CO₂ discharges or reproduce the whole data-set, respectively. Thus, the rock layer appears the most representative radon source, controlling radon production and whose thickness constrains the degassing depth at both sites. The degassing depth appears nevertheless better constrained at SBHS, where depths greater than a few hundreds of metres are unlikely, whereas at FLFF, depths of ≈ 10 km might be possible (Supp. Figs. S8 and S11). Our model clearly interprets the unambiguous observation of the saturation of the radon flux at high CO₂ flux as a smaller source thickness. At high CO₂ velocities, the path length in the rock is not sufficient to recharge the CO₂ flow with radon.

Our model takes into account the temperature dependence of radon source term and other radon parameters, but only experimentally constrained in the range 0–100 °C (see “Methods” section), highlighting also limitations of our approach. Too little information on radon parameters variation with higher temperature is available⁶⁴, motivating future investigations in that direction. Similarly, at such degassing depths (hundreds to thousands of metres), confining pressure is high, but, despite few studies on granitic¹³ and volcanic rocks¹², radon source term variation with increasing pressure remains poorly known. Nevertheless, our model indicates that the variations of the source parameters or their heterogeneity are second-order effects. The largely dominating effect remains in all reasonable instances the path length, and hence the source depth.

The combination of field measurements of coupled CO₂ and radon fluxes at the surface, laboratory characterisation of radon source term, and radon transport modelling have constrained the depth of CO₂ degassing at two hydrothermal sites in different tectonic contexts and with different CO₂ sources. Our results thus attest to the relevance of gas flux monitoring, particularly important at sites under near-critical conditions, sensitive to Earth's deformation and earthquakes, which, contrary to what has been done in the past^{35,65}, are not exclusively found in volcanic regions. For example, the $M_w 7.9$ 2015 Gorkha earthquake greatly affected gas emissions at SBHS, likely liberating CO₂ previously stored in a crustal reservoir at shallow depths⁸. Our combined approach shows that emitted radon may have changed from a deeper (≈ 1000 m) to a shallower source (≈ 100 m) following the earthquake (Supp. Fig. S12).

Our combined study has revealed to be a powerful tool to determine the depth of degassing at FLFF, with mantle and magmatic CO₂ sources, and at SBHS, with a metamorphic CO₂ source. To date, only few data are reported on CO₂ and radon fluxes together⁴⁷, almost all of them in collision context. Our approach should be systematically applied to sites in other tectonic contexts, such as rifting, reverse fault, strike-slip fault, or subduction, as well as in other volcanic environments. In addition, our model was found particularly sensitive for CO₂ flux > 5000 g m⁻² day⁻¹, motivating future improvement in the measurement of such high fluxes. The presence of high gas fluxes, especially CO₂, will be important to investigate in more details in the future, in particular when re-evaluating global carbon budgets^{9,66,67}.

The radioactive gas radon, tracking CO₂ degassing and diagnosing CO₂ transport mechanisms, emerges as a powerful asset to characterise gaseous emissions and monitor earthquake-sensitive geosystems. In addition, the exceptionally high radon discharge from FLFF (≈ 9 GBq day⁻¹), during a quiescent period, raises the issue of unconstrained radon emission from volcanoes, and suggests significant radon output from volcanic areas worldwide, especially when a large eruption occurs. Because emissions from volcanoes, unlike background diffusive soil emission, can reach above the atmospheric boundary layer, radon release from volcanoes worldwide may have substantial effects on atmosphere ionisation, aerosol formation, and climate⁶⁸.

Methods

Radon flux. Radon-222 flux was measured at the surface using the accumulation chamber method^{46,56}. Increase rate with time of radon activity concentration inside the chamber, directly related to radon flux, was measured using scintillation flasks (Alcade, France) at both sites. Radon concentration in the flasks was inferred 3.5 h after sampling from counting in photomultipliers (CALEN™, Alcade, France), regularly inter-calibrated in the laboratory. The method is robust, even in remote location⁴⁸, and reliable where radon fluxes range over several orders of magnitude⁵⁴. Radon flux is expressed in 10⁻³ Bq m⁻² s⁻¹. Associated uncertainty (Supp. Fig. S1a) was estimated from several systematic tests⁵⁴. Relative experimental uncertainty ranges from 15% for fluxes $\approx 100 \times 10^{-3}$ Bq m⁻² s⁻¹ to 30% for fluxes $\approx 10 \times 10^{-3}$ and $\approx 10,000 \times 10^{-3}$ Bq m⁻² s⁻¹. Radon fluxes were measured during stable weather conditions, in summer 2016 at FLFF, and in winters 2009–2011, 2015, and 2016 at SBHS. A majority of points were measured several times (from 2 to 18 times); point-averages are arithmetic means. At FLFF, based on several measurements along time at selected points, variations of radon flux were $< 10\%$ and $< 15\%$ for fluxes $< 1000 \times 10^{-3}$ and $> 10,000 \times 10^{-3}$ Bq m⁻² s⁻¹, respectively. At SBHS, temporal variations of radon flux were generally $< 30\%$ for all fluxes. Total radon discharge, expressed in MBq day⁻¹ (or Bq s⁻¹), was estimated following the sequential Gaussian simulations (sGs) method⁶⁹ with 100 equiprobable realizations. Radon fluxes measured at SBHS (gas zones 1 and 2) were published^{18,47,54} (Table 1).

CO₂ flux. CO₂ flux was measured at the surface using the accumulation chamber method⁷⁰. Increase rate with time of CO₂ concentration inside the chamber, directly related to CO₂ flux, was measured by several portable sensors: at FLFF, two portable infrared fluxmeters (WestSystem™ CO₂Flux, Italy), regularly calibrated by the manufacturer and inter-calibrated in the field; at SBHS, two portable infrared CO₂ sensors and home-made accumulation chambers (Testo™ 535, Testo AG, Germany; Vaisala™ CARBOCAP® Hand-Held GM70, Finland), regularly inter-calibrated in the laboratory. Portable instruments are robust, even in remote location⁴⁸ or during monsoon⁵⁶, and reliable where CO₂ fluxes range over six orders of magnitude^{32,54}. CO₂ flux is expressed in g m⁻² day⁻¹. Associated uncertainty (Supp. Fig. S1b) was estimated from previous assessments⁷¹ and systematic tests⁵⁴. Relative experimental uncertainty ranges from 10% for low fluxes (10 g m⁻² day⁻¹) to 30% for large fluxes (10,000 g m⁻² day⁻¹). CO₂ fluxes were measured during stable weather conditions, in summer 2016 at FLFF, and in winters 2015 to 2020 at SBHS. Every point was measured several times (from 2 to 26 times); point-averages are

arithmetic means. At FLFF, only 4.3% temporal variation of CO_2 flux was recorded at the 1-h continuous fluxmeter station GFUR2 during the 8-day-long field campaign (mean: $268 \pm 12 \text{ g m}^{-2} \text{ day}^{-1}$). At SBHS, several points measured along time during the numerous field campaigns from November 2015 to January 2020 showed variations $< 10\%$ for fluxes $< 100 \text{ g m}^{-2} \text{ day}^{-1}$ and $< 20\%$ for fluxes $> 1000 \text{ g m}^{-2} \text{ day}^{-1}$. Total CO_2 discharge, expressed in t day^{-1} (or mol s^{-1}), was estimated following the sGs method⁶⁹ with 100 equiprobable realizations. CO_2 fluxes measured from November 2015 to January 2018 at SBHS (gas zones 1 and 2) were published⁸ (Table 1).

Radon– CO_2 fluxes correlation. Only values of radon and CO_2 fluxes obtained at the same measurement points were considered and no interpolation was used. This represents a total of 136 and 157 locations of combined fluxes for FLFF and SBHS, respectively.

Radon sources. *Effective radium-226 concentration (EC_{Ra}).* Radon-222 source in porous materials is the effective radium-226 concentration, expressed in Bq kg^{-1} , i.e. the product of the bulk radium concentration C_{Ra} and the emanation coefficient E , probability that a radium atom produces a radon atom in the pore space. EC_{Ra} was measured in the laboratory on rock and soil samples using a radon accumulation method^{60,72,73}. The sample was placed in a hermetically closed container. After a 5-to-18-day accumulation time, radon concentration of the free air inside the accumulator was determined after sampling using a scintillation flask (Algade, France) and counting in a photomultiplier (CALEN™, Algade, France). A minimum of three measured values per sample were averaged. Relative experimental uncertainty ranges from 3% for $EC_{\text{Ra}} \approx 30 \text{ Bq kg}^{-1}$ to 10% for $EC_{\text{Ra}} \approx 1 \text{ Bq kg}^{-1}$. SBHS data were partly published^{50,54}.

Ground radon concentration. Ground radon-222 concentration, expressed in Bq m^{-3} , was measured at FLFF at 60–100 cm depth by gas pumping using a portable RAD7 detector (DurrIDGE Company, Inc., USA). At SBHS, it was measured at 100 cm depth using continuous radon concentration probes (Barasol™ and BMC2™, Algade, France). All these instruments are based on the detection of alpha particles by a silicon detector material. Sensitivity, inter-comparison dispersion, and overall common uncertainty are 4 and 50 Bq m^{-3} , 5% and 3%, and 5% and 5%, for RAD7 and continuous instruments, respectively. SBHS data were partly published^{54,57}.

Radon in water bubbles. Radon-222 concentration in water bubbles was measured after accumulation of gas in small containers above the water pond, sampling using scintillation flasks, and counting in photomultipliers. Relative experimental uncertainty is similar to that of EC_{Ra} method.

Radon and radium in water. Radon-222 concentration in water, expressed in Bq L^{-1} , was measured by the emanometry method^{58,62}. The water was sampled in a container, hermetically closed after sampling. After shaking, sampling of the air inside the container was done by a scintillation flask and counting using a photomultiplier. Relative experimental uncertainty ranges from 5 to 30%. Two or more replicates were performed at each location. Radium-226 concentration in water, expressed in $10^{-3} \text{ Bq L}^{-1}$, was measured following the radon method^{58,62,74}. After water sampling in the field, the container was kept closed in the laboratory for 2 months minimum before measurement. Relative experimental uncertainty ranges from 5% for $C_{\text{Ra}} \approx 100 \times 10^{-3} \text{ Bq L}^{-1}$ to 10% for $C_{\text{Ra}} \approx 10 \times 10^{-3} \text{ Bq L}^{-1}$ (Table 2).

CO_2 sources. *Ground CO_2 concentration.* Ground CO_2 concentration, expressed in vol.%, was measured at FLFF at 60–100 cm depth by gas pumping using a portable infrared sensor (Geotechnical Instruments, UK) with 0–100 vol% measurement range. At SBHS, gas was sampled using evacuated glass tubes and CO_2 concentration was determined manometrically⁵⁴. Absolute experimental uncertainty ranges from 0.5 to 3% for the infrared cell and from 0.1 to 1% using the mass spectrometer. Only 2007–2018 SBHS data were partly published^{8,46,47,54}.

$\delta^{13}\text{C}$ of CO_2 gas. The gas was collected using accumulation chambers and evacuated glass tubes^{8,47,54}. Gas samples were analysed for molecular composition in the laboratory⁴⁵. Carbon isotopic ratio of gaseous CO_2 , $\delta^{13}\text{C}$, expressed in ‰ and defined relative to the standard value of Pee Dee belemnite (V-PDB), was determined after off-line purification using a Finnigan™ MAT-253 mass spectrometer (Thermo Electron Corp., Germany) at CRPG (Nancy, France). Repeatability and absolute experimental uncertainty are 0.1 ‰. FLFF data were partly published^{7,22}; SBHS 2007–2018 data were published^{8,46,47,54}.

Dissolved inorganic carbon concentration and $\delta^{13}\text{C}$ of dissolved CO_2 . Water was sampled using 12-mL screw cap vials. Dissolved inorganic carbon (DIC) concentration, expressed in $10^{-3} \text{ mol L}^{-1}$, was determined using a gas chromatograph coupled to an isotope ratio mass spectrometer (GCIRMS, GV 2003, GV Instruments, UK) at IPGP (Paris, France). Relative experimental uncertainty ranges from 1 to 2%. $\delta^{13}\text{C}$ of dissolved CO_2 in water, expressed in ‰ relative to V-PDB, was determined using the same spectrometer at IPGP. Absolute experimental uncertainty is 0.1‰. SBHS data were published⁸; FLFF data were taken from the literature²⁴ (Table 2).

Surface and ground temperature. *Surface temperature.* At FLFF, temperature was measured systematically at surface using a thermocouple (Testo 925™, Testo AG, Germany). Absolute experimental uncertainty is 0.1 °C. At SBHS, temperature was measured at the surface using regularly inter-calibrated, autonomous sensors (SB39, Seabird™, USA). Their sensitivity is better than $3.5 \times 10^{-3} \text{ °C}^{54}$.

Ground temperature. At FLFF, temperature was measured at the top of 60–100 cm holes with the same thermocouple used for surface temperature. Absolute experimental uncertainty is 0.1 °C. At SBHS, temperature was measured at 100 cm depth using SB39 sensors, recording at 30 s intervals (Table 2).

Radon transport modelling. *Diffusive–advective radon transport model.* The model considers that degassing of hydrothermal CO₂ is initiated at depth, provided it travels fast enough relative to the radon-222 half-life. After degassing, CO₂ percolates through rock and soil layers until it reaches the surface, carrying radon produced by these layers along its pathway. We consider soil, rock, and deep rock layers from surface in a semi-infinite porous medium (Supp. Fig. S6). For radon, we consider gaseous, dissolved, and adsorbed phases with equation terms of diffusion, advection, radioactive decay, and production, that are implemented analytically. We use transport Eqs. (B2)–(B4) of Ref.⁵⁸ without modification. Using Eqs. (B6)–(B11) or Eqs. (B13)–(B17) for large advection (Ref.⁵⁸), we solve the following steady-state 1D equation for each layer i :

$$D_{Si} \frac{\partial^2 C_{ai}}{\partial z^2} - T_{Si} \frac{\partial C_{ai}}{\partial z} + \lambda(C_{ai}^0 - C_{ai}) = 0,$$

where $D_{Si} = \frac{\varepsilon_{ai} D_{ei}^a + \varepsilon_{wi} \kappa_{wi} D_{ei}^w}{\varepsilon_{ai} + \varepsilon_{wi} \kappa_{wi} + k_{di} \rho}$, $T_{Si} = \frac{u + I \kappa_{wi}}{\varepsilon_{ai} + \varepsilon_{wi} \kappa_{wi} + k_{di} \rho}$ and equilibrium radon concentration $C_{ai}^0 = \frac{EC_{Ra1} \rho}{\varepsilon_{ai} + \varepsilon_{wi} \kappa_{wi} + k_{di} \rho}$, with u the Darcy velocity related to the carrier CO₂ flux and I the water infiltration (for more details see Supp. Fig. S6 and Ref.⁵⁸). Here, for given layer i , we introduce empirical relations for porosity- and water-saturation-dependent effective diffusion coefficients⁷⁵ ($D_{ei}^a = 1.1 \times 10^{-5} \times \varepsilon_{ai} \exp(-6S_{wi} \varepsilon_{ai} - 6S_{wi}^{14} \varepsilon_{ai})$) and $D_{ei}^w = 1.1 \times 10^{-8} \times \varepsilon_{wi} \exp(-6(1 - S_{wi}) \varepsilon_{wi} - 6(1 - S_{wi})^{14} \varepsilon_{wi})$), temperature-dependent water/air partition⁵⁸ and adsorption coefficients⁷⁶ ($\kappa_{wi} = 0.104 + 0.416 \exp(-0.0491 T_i)$) and $k_{di} = k_{di}^0 \exp\left(\frac{19500}{8.314} \left(\frac{1}{T_i} - \frac{1}{253.15}\right)\right)$ with $k_{d1}^0 = 0.181$ and $k_{d2,3}^0 = 0.258$) and water-saturation- and temperature-dependent radon source terms^{73,77} ($EC_{Ra1} = EC_{Ra1}^0 (32.22 \exp(-1.88 S_{w1}) - 31.43 \exp(-1.98 S_{w1})) (1 + 0.01 (0.7875 T_1 - 15.75))$) and $EC_{Ra2,3} = EC_{Ra2,3}^0 (32.22 \exp(-1.88 S_{w1}) - 31.43 \exp(-1.98 S_{w1})) (1 + 0.01 (0.5535 T_1 - 11.07))$). For both sites, we fix parameters according to field and laboratory data-sets, such as soil thickness (1 m), soil and rock porosity and water saturation (deep rock is assumed saturated), gas temperature, and radon source term; we calculate water–air partition and adsorption coefficients; and we correct radon source term. For a given CO₂ flux, we calculate a radon flux using Eqs. (B12) or (B18) for large advection (Ref.⁵⁸). We then compare the calculated and measured radon fluxes using a χ^2 coefficient (see text) to estimate the best depth of degassing (interface between the rock and deep rock layers). We separately tested the sensitivity of the degassing depth to several parameters considering an acceptable range of values (Supp. Fig. S7). For the soil and deep rock layers, varying total porosity (soil: 0.01–0.4, rock: 0.01–0.15), water saturation (0.1–0.9), gas temperature and radon source term (50% around the mean; see Table 2) has low effect on the degassing depth. However, such variations for the rock layer yield larger effects (Supp. Fig. S7), in particular for gas temperature and radon source term. Then, to explore the phase space, similarly to a Bayesian Monte Carlo unbiased approach, we generated 500 independent models, varying gas temperature and radon source term in the soil and rock layers, with Gaussian distributions, to estimate the depth of degassing (see text; Table 3). The final degassing depth estimate is taken as the median value from the 500 simulations (Supp. Fig. S8).

Alternative models. Other modelling approaches were also used. A first analytical model⁵⁸ considers that a given fraction f of CO₂ degasses together with radon from water springs at or near the surface. We use Eqs. (A1)–(A7) of Ref.⁵⁸ without modification. Another analytical model⁵⁸ considers that radon transport is only governed by diffusion, using the same equations given above with $u = 0$. We use Appendix C of Ref.⁵⁸. For these two alternative models, the full description and the equations are given in a previous contribution⁵⁸.

Data availability

The datasets generated and analysed during the current study are available from the corresponding author on reasonable request.

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Author contributions

F.G., F.V. and S.B. designed this work. F.G., F.V., and F.P. wrote the paper. For the FLFF data-set: F.G. measured radon and CO₂ fluxes, dissolved radon concentration in water, EC_{Ra} in the laboratory, and collected gas and water samples; F.V. measured CO₂ flux and ground temperature; C.S. measured ground radon and CO₂ concentrations and temperature; J.E.P. measured radon and CO₂ fluxes; F.V., C.S., and J.E.P. collected soil and rock samples; C.F.-L. measured carbon isotopic composition of the gas phase in the laboratory; F.P. measured EC_{Ra} and dissolved radium concentration in the laboratory. For the SBHS data-set: F.G. measured radon and CO₂ fluxes, dissolved radon concentration in water, EC_{Ra} in the laboratory, and collected gas, water, soil, and rock samples; S.T. measured CO₂ fluxes and collected water in 2020; L.B.A., B.P.K., and M.B. participated to field CO₂ flux measurements; P.A. measured carbon isotopic composition and dissolved inorganic concentration of the water phase in the laboratory. C.F.-L. measured carbon isotopic composition of the gas phase in the laboratory; F.P. measured radon fluxes, dissolved radon and radium concentrations in water, and EC_{Ra} in the laboratory. F.G. processed Figs. 2, 5, and Figs. S1–S10. F.V. processed Figs. 1a, 3 and 4. S.T. processed Fig. 1b. F.G. and F.P. did

the modelling. V.Z., J.V., and S.B. discussed the modelling. All the authors discussed the data-sets and amended the manuscript.

Competing interests

The authors declare no competing interests.

Additional information

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