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Abiotic carbonate dissolution traps carbon in a semiarid desert

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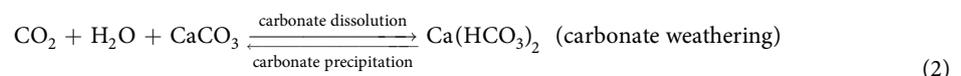
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It is generally considered that desert ecosystems release CO₂ to the atmosphere, but recent studies in drylands have shown that the soil can absorb CO₂ abiotically. However, the mechanisms and exact location of abiotic carbon absorption remain unclear. Here, we used soil sterilization, ¹³CO₂ addition, and detection methods to trace ¹³C in the soil of the Mu Us Desert, northern China. After ¹³CO₂ addition, a large amount of ¹³CO₂ was absorbed by the sterilised soil, and ¹³C was found enriched both in the soil gaseous phase and dissolved inorganic carbon (DIC). Further analysis indicated that about 79.45% of the total ¹³C absorbed by the soil was trapped in DIC, while the amount of ¹³C in the soil gaseous phase accounted for only 0.22% of the total absorbed ¹³C. However, about 20.33% of the total absorbed ¹³C remained undetected. Our results suggest that carbonate dissolution might occur predominately, and the soil liquid phase might trap the majority of abiotically absorbed carbon. It is possible that the trapped carbon in the soil liquid phase leaches into the groundwater; however, further studies are required to support this hypothesis.

Drylands (arid and semiarid desert ecosystems) cover about 41% of the global terrestrial surface¹ and release a large amount of carbon via soil CO₂ flux^{2,3}. However, a net uptake of carbon is observed in some desert ecosystems, especially during nighttime, such as the Mojave Desert^{4,5}, a playa of the Great Basin Desert⁶ and the Chihuahuan Desert⁷ in North America, the Gurbantunggut Desert^{8,9} and the Mu Us Desert^{10,11} in Asia and a dry valley in Antarctica¹². Although the carbon absorptions in these cases are various, it may also play an important role in the terrestrial carbon cycle because of the great areal extent of drylands^{10,13}. The ‘anomalous’ and variable soil carbon accretion suggests that the soil carbon cycle in desert ecosystems is quite complicated^{14,15}.

Although it is known that CO₂ absorption by the soil is induced by abiotic processes and is relevant to some ambient factors (e.g. temperature, soil water content, and air pressure)^{8,10,12,16,17}, previous studies failed to reveal the underlying mechanism of the abiotically absorbed CO₂ by the soil, which is essential to better understand CO₂ absorption by the soil in drylands¹⁸. Applying the ¹³C isotope tracer method to undisturbed soil, we previously found that the majority of absorbed carbon might conserve in the soil solid phase¹⁹; however, we could not demonstrate the mechanism of abiotic CO₂ absorption and sequestration because of the influence of biotic processes. Hence, further studies on the abiotically absorbed carbon is required.

It is surmised that abiotic carbonate dynamic is a major contributor of CO₂ absorption by the soil^{10,20} and the involved abiotic processes can be expressed as²¹:



Since Ca-silicate weathering is an extremely slow process, it can be ignored at the diel or annual scale²², while the process of carbonate weathering may help to better explain CO₂ absorption by the soil. However, the reactions are reversible. Whether carbonate dissolution can predominately occur and induce the abiotic atmospheric CO₂ absorption by soil in drylands, have not been proven directly by any previous study currently. Here, we hypothesized that carbonate dissolution can induce the abiotic atmospheric CO₂ absorption by desert soils. To test the hypothesis, we used the sterilization, leaching method and ¹³CO₂ isotope tracer technique to investigate

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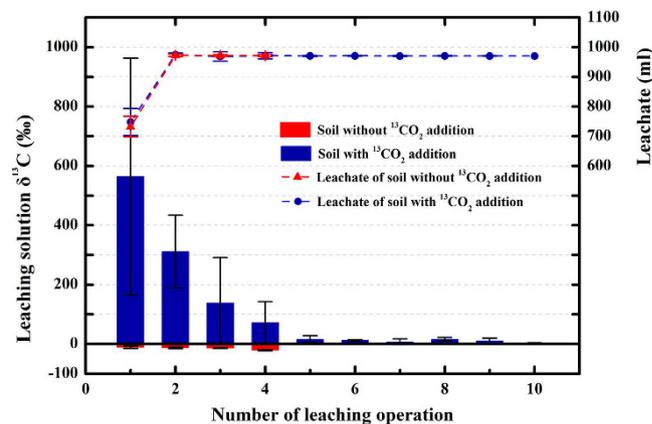


Figure 1. ¹³C abundances (δ¹³C; ‰) of leaching solution of sterilised soil with and without ¹³CO₂ addition, and the amount (ml) of the corresponding leaching solution. Blue columns and the blue dash line refer to sterilised soil with ¹³CO₂ addition. Red columns and the red dash line refer to sterilised soil without ¹³CO₂ addition. Error bars represent standard error of the mean.

the abiotic soil CO₂ exchange and ¹³C abundance of soil liquid and vapour phase in the desert soil of the Mu Us Desert, northern China.

Results

The amount of abiotic ¹³CO₂ absorption. After ¹²CO₂ was replaced in the chamber, the initial ¹³CO₂ amount in the atmosphere of the chamber was 446 μmol. After 12 h (throughout the nighttime), the amount of the final atmospheric ¹³CO₂ in the chamber was 149 μmol. The amount of ¹³CO₂ absorption by the sterilised soil reached to 297 μmol.

¹³C abundance (δ¹³C) in the leaching solution with and without ¹³CO₂ addition. The ¹³C abundances and amounts in the leaching solution that removed by frequent leaching are illustrated in Fig. 1. The amount of the collected leaching solution of sterilised soil with ¹³CO₂ addition did not differ remarkably from that without ¹³CO₂ addition at each leaching operation, and both approximately equal to 1,000 ml since the second leaching operation. However, the leachate δ¹³C (δ¹³C of DIC in leachate) of sterilised soil with ¹³CO₂ addition showed intense enrichment (although leachate δ¹³C decreased with the increasing number of leaching operations, all the values of leachate δ¹³C were positive, ranging from 564.48 ± 399.21% to 2.85 ± 1.31%), compared with the leachate δ¹³C of sterilised soil without ¹³CO₂ addition (all the values of leachate δ¹³C were negative, ranging from -11.40 ± 3.64% to -21.21 ± 1.89%).

¹³C abundance (δ¹³C) in the soil gaseous phase with and without ¹³CO₂ addition. The abundances of ¹³C of soil air with and without ¹³CO₂ addition are showed in Fig. 2. The δ¹³C of soil air without ¹³CO₂ addition was -9.3 ± 2.26% after soil-atmosphere CO₂ exchange in the chamber. While the δ¹³C of soil air with ¹³CO₂ addition was enriched after the CO₂ exchange in the chamber (599.2 ± 192.87%).

Absorbed ¹³CO₂ tracing. The total ¹³C in the soil leaching solution was calculated at about 235.98 μmol, and the ¹³C in the soil air was 0.66 μmol. Thus, the accumulated amount of absorbed ¹³C in the soil liquid phase accounted for 79.45% of the total absorbed ¹³C, while the amount in the soil gaseous phase accounted for only 0.22%. In addition, about 20.33% of the total absorbed ¹³C remained undetected.

Discussion

79.45% of the labelled carbon (¹³CO₂) was found in dissolved inorganic carbon (H¹³CO₃⁻), indicating that the majority of CO₂ absorbed abiotically by the soil during nighttime is trapped in the soil liquid phase and converted into dissolved inorganic carbon (DIC) at this study site. A recent study also showed that the soil liquid phase in a saline/alkaline desert contains amounts of newly formed DIC, and the formation of DIC may be associated with atmospheric CO₂²³. However, in our previous study, it was found that the majority of absorbed ¹³C was fixed in the solid phase of undisturbed soil¹⁹. The discrepancy between the two studies could be attributed to the influence of biotic processes. Verrecchia and Verrecchia²⁴ reported that bacteria and fungi could accelerate the formation of needle fibre calcite (CaCO₃). The impact of rhizosphere processes may also be important in the formation of carbonate. For instance, the roots absorb soil water and can cause bicarbonate decomposition producing carbonate²⁵. In this study, the effect of biotic processes was excluded by soil sterilisation; therefore, the carbon absorbed by the soil could be only trapped abiotically. Moreover, soil water participation and carbon transportation may be another reason for the discrepancy. These may be overlooked in the previous study. Ignoring the interference of biotic processes, conserving in the soil solid phase may be a medium status for the carbon absorbed by soil from atmosphere according to the results in this study. Therefore, this work is an extension of the previous one.

Variations in the ambient temperature can affect in DIC formation and accumulation^{26–28}. The high temperature results in DIC decomposition²⁹, while the low temperature induces DIC accumulation¹². In this study, the

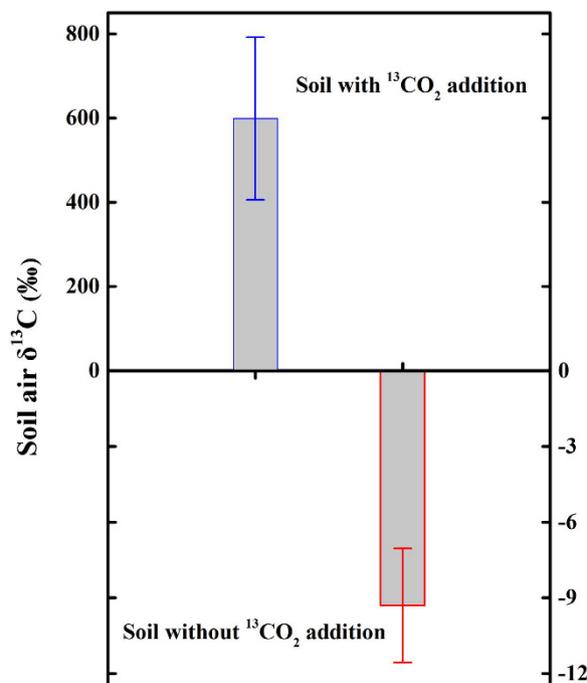


Figure 2. ^{13}C abundances ($\delta^{13}\text{C}$; ‰) of soil air with and without $^{13}\text{CO}_2$ addition. Grey column with blue border refers to sterilised soil with $^{13}\text{CO}_2$ addition. Grey column with red border refers to sterilised soil without $^{13}\text{CO}_2$ addition. Error bars represent standard error of the mean.

ambient temperature was low during nighttime. As a result, the process of carbonate dissolution may predominantly occur in the soil. In subsoil, the temperature may be persistently lower than topsoil (i.e. the sampled soil in this study)³⁰. More DIC may be formed and accumulated without decomposition. Previous studies showed that the residence time of carbon in pedogenic DIC could be over three orders of magnitude higher than that of soil organic matter^{31–34}. Therefore, the newly formed DIC may not easily turn into CaCO_3 abiotically until the ambience is changed. The accumulation of DIC may induce a continuous consumption of soil CO_2 ; inorganic consumption of soil CO_2 through carbonate dissolution can create a pressure gradient of CO_2 between the soil and the atmosphere, and then the pressure gradient can induce atmospheric CO_2 to be pumped into the soil to replenish soil CO_2 (as the ingredient in carbonate dissolution)¹⁰.

Although the results showed that the soil could absorb CO_2 from the atmosphere and then trap large amounts of the atmospheric carbon by forming and accumulating DIC at this study site, we still cannot declare that this is an important soil carbon sink, because we have no direct evidence to support that DIC is sequestered in the soil at a long-term scale. For the formation of soil carbonate induced by biotic processes, CO_2 , as a by-product, releases back to the atmosphere³⁴ and overshadows the abiotic CO_2 absorption³⁵. As a result, carbonate in the soil solid phase may not be the main destination of absorbed atmospheric CO_2 at the diel or annual scale. The hydrosphere beneath the soil may be a potential outlet of trapped carbon. Figure 1 shows that the vertical migration of soil water occurs easily, and the majority of the absorbed ^{13}C in DIC can be washed out by four leaching operations. These results suggested that DIC may finally transport to the aquifer, regardless of being a long-term process. Similarly, Ma *et al.*¹⁸ and Li *et al.*²³ reported that soil DIC could be leached into the aquifer by rainfall or snow glacier melt and irrigation water or river water in arid deserts. Moreover, Walvoord *et al.*³⁶ found that the accumulated nitrate in the subsoil might also transport into the groundwater because of long-term leaching in xeric ecosystems. In the study site, groundwater is recharged mainly by rainfall³⁷ and has almost no hydrological connection to rivers and oceans. If the trapped carbon transports deeply into the aquifer, the low and constant temperature and alkalinity of groundwater may convert the aquifer into a reservoir of carbon beneath the desert soil. However, further studies are required to test whether the newly formed DIC can be leached from the soil to the hydrosphere.

Although this is a case study in the Mu Us Desert, the results can provide the direct and effective support to the hypothesis that carbonate dissolution can induce the abiotic atmospheric CO_2 absorption by desert soils. It is undeniable that the amount of carbon trapped in DIC may be overestimated, because the deionized water used for leaching can inevitably dissolve some of the solid fraction (soil carbonate) in theory. To diminish the overestimation, we multiply leached with much less deionized water each time rather than one time leaching with the large amount of deionized water (the duration of each leaching operation is much shorter and the amount of deionized water used for leaching is much less each time). Therefore, the overestimation may be small, and its influence should be slight.

After $^{13}\text{CO}_2$ addition and exchange in the chambers for 12 h, ^{13}C in the soil gaseous phase was enriched (Fig. 2); however, the amount of newly conserved ^{13}C in the soil gaseous phase was much less than that fixed by the soil liquid phase (accounted for about 0.28% of the fixed ^{13}C in the soil liquid phase). These results were in

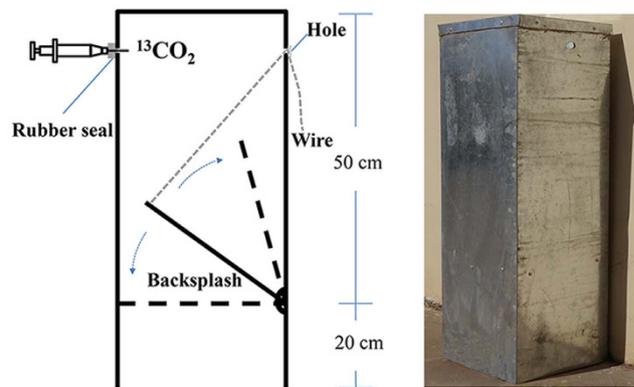


Figure 3. Structure of the $^{13}\text{CO}_2$ exchange chamber. Neutral glass cement was used to seal the hole immediately after the backslash was opened or closed during the process of $^{13}\text{CO}_2$ addition to ensure the chamber sealability.

agreement with those reported in our previous study¹⁹, suggesting that the soil gaseous phase may only serve as the connecting medium between the atmosphere and the soil liquid phase. CO_2 in the soil gaseous phase may easily transport upwards to the atmosphere or downwards to the alkaline soil solution.

About 20.33% of the total absorbed ^{13}C remained undetected, probably because it was stored in the soil solid phase. Since the leaching operations started at 9:00 A.M., soil temperature might already begin to rise, and a part of DIC probably turned into CaCO_3 and released as CO_2 . The results from abiotic soil CO_2 flux measurement showed that soil could absorb atmospheric CO_2 during nighttime and release CO_2 during daytime, somewhat supporting our hypothesis. It is notable that even though DIC can turn into CaCO_3 during daytime, the amount of lapsed DIC may be small, because the abiotic net carbon exchange between the soil and the atmosphere is usually negative at the diel scale (negative net carbon exchange represents carbon absorption by soil)¹¹. Therefore, at the diel scale, most of the abiotically absorbed carbon may also be trapped in the soil liquid phase.

Methods

Site description. The study site is located on the southwestern fringe of the Mu Us Desert ($37^\circ 42' \text{N}$, $107^\circ 13' \text{E}$; 1,509 m above sea level), northern China. It has a temperate continental monsoon climate with a mean annual temperature of 7.6°C , mean annual solar radiation of $1.4 \times 10^5 \text{ J cm}^{-2}$, and mean annual wind speed of 3 m s^{-1} (prevailing northwest wind). The frost-free period lasts around 128 d. The mean annual precipitation is 275 mm (1954–2013), mainly occurring in August and September³⁵. The soil type is Aripsamment (derived from aeolian sand; soils with high CaCO_3 content tend to be salinised). The soil (0–20-cm depth) comprises 94.8% sand, 4.5% silt, and 0.7% clay and has a pH of 8.6¹⁰. Soil bulk density is 1.54 g cm^{-3} and soil porosity 42%¹⁹. The study site is sparsely vegetated by *Artemisia ordosica*, *Astragalus mongolicum*, *Salix psammophila*, and *Tamarix chinensis* (canopy coverage $\leq 30\%$).

$^{13}\text{CO}_2$ tracing. To test whether carbonate dissolution occurs and detect the location of the abiotically absorbed carbon in the soil, an improved leaching operation (adopted in Li *et al.*²³; for extraction DIC) and $^{13}\text{CO}_2$ tracing experiment (used in Liu *et al.*¹⁹; for carbon tracing) were carried out in September and October 2014 using iron $^{13}\text{CO}_2$ exchange chambers (25 cm in length; 25 cm in width; 70 cm in height; 0.1 cm in thickness; Fig. 3). Three quadrate steel soil collars (25 cm in length; 25 cm in width; 20 cm in height; 0.2 cm in thickness) were placed randomly in the study site with a 2-cm wall to be exposed above the soil surface in order to allow the installation of the iron $^{13}\text{CO}_2$ exchange chambers. The soil within the collars was equilibrated with its surrounding for 24 h to minimise the disturbing effect. Subsequently, soil samples were collected and sterilised as described by Xie *et al.*⁸. The sterilised soil was placed *in situ* and equilibrated with its surrounding for 17 h to minimise the disturbing effect.

Each $^{13}\text{CO}_2$ exchange chamber (with backslash opened) was immersed in NaOH solution (5 mol L^{-1}) up to about 2 cm for 2 h to remove $^{12}\text{CO}_2$. Then, two $^{13}\text{CO}_2$ exchange chambers (with backslash closed) were installed onto the collars (pushed 15 cm deep into the soil), while the third collar was used as a control. We injected 10 ml $^{13}\text{CO}_2$ (concentration $> 99.99\%$) into each chamber, opened the backslash, and allowed the soil to exchange $^{13}\text{CO}_2$ for 12 h.

The backslash of each chamber was closed after the $^{13}\text{CO}_2$ exchange, the gaseous samples (140 ml) in each chamber were collected with an aluminium foil gas-collecting bag, and the $^{13}\text{CO}_2$ exchange chambers were removed. In order to increase accuracy, a polyvinyl chloride (PVC) sheet (25 cm in length; 25 cm in width; 25 cm in height; 0.1 cm in thickness) was used to divide the sterilised soil into two parts; one for soil CO_2 extraction (25 cm in length; 5 cm in width; 18 cm in height) and the other for leaching (25 cm in length; 20 cm in width; 18 cm in height). An aluminium foil gas-collecting bag (200 ml) was used to reserve the extracted soil air (140 ml) from the one part of sterilised soil, and a PVC cylinder (15.3 cm outer diameter; 15 cm inner diameter; 25 cm in height) was inserted into the other part of sterilised soil to sample the soil column for leaching. The cylinder was sealed at the top and bottom with base plates and transported to the laboratory.

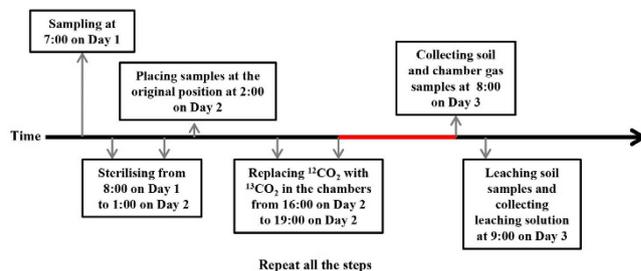


Figure 4. Flow diagram of $^{13}\text{CO}_2$ tracing. Red line indicates CO_2 exchange in the chamber.

A layer of gauze (100-mesh) was placed under the bottom plate to prevent mud from leaching. Then, the bottom plate was removed, and the sterilised soil cylinder was leached with 1,000 ml deionised water. The sterilised soil with $^{13}\text{CO}_2$ addition was leached 10 times (about 10 times of leaching may leach out almost all the ^{13}C according to the study of Ma *et al.*¹⁸), while the control (without $^{13}\text{CO}_2$ addition) 4 times. We assumed that the abundance of ^{13}C would not change considerably, because the natural ^{13}C abundance of the soil was low³⁸, and thus, the different number of leaching operations could not influence the comparison between the results from sterilised soil with and without $^{13}\text{CO}_2$ addition. After each leaching operation, we weighed the total volume of leaching solution (V_{leachate}) and collected leaching solution samples of 30 ml using brown glass bottles. During all these processes, the ambient temperature was below 10°C .

The number of replications for the sterilised soil with and without $^{13}\text{CO}_2$ addition was 18 and 9, respectively. The flow of $^{13}\text{CO}_2$ tracing is shown in Fig. 4. The $\delta^{13}\text{C}$ and CO_2 concentration of all gaseous samples were measured by a carbon dioxide isotope analyser (CCIA-EP 912-0003; Los Gatos Research, Mountain View, CA, USA) and the $\delta^{13}\text{C}$ of all liquid samples ($\delta^{13}\text{C}$ of DIC in the soil liquid phase) by an isotope mass spectrum analyser (Finnigan MAT253 Gas Bench-IRMS; Thermo Fisher Scientific Inc., Waltham, MA, USA) as described by Liu *et al.*¹⁹.

Measurements of soil volumetric water content (VWC) and bicarbonate radical concentration (C_{DIC}). VWC was monitored from 20:00 to 7:00 the following day at a depth of 10 cm using the ECH2O system (LI-COR, Lincoln, NE, USA) with five Em50R sensors placed near the soil collars, and data were logged every 1 h. After collecting liquid samples (without $^{13}\text{CO}_2$ addition), C_{DIC} was measured using the conventional method of acid base titration³⁹.

Data processing and analysis. The absorbed $^{13}\text{CO}_2$ within the soil collar was calculated as follows:

$$N = \frac{1000 \times V_{\text{initial}}}{R_t} - \frac{C_{\text{final}} \times V_{\text{chamber}}}{R_t}, \quad (3)$$

where N is the total absorbed $^{13}\text{CO}_2$ (μmol), V_{initial} is the initial volume of $^{13}\text{CO}_2$ (10 ml), C_{final} is the final $^{13}\text{CO}_2$ concentration ($\mu\text{mol mol}^{-1}$), V_{chamber} is the volume of the chamber exposed above the soil surface (L), and R_t is the molar volume of gas (L mol^{-1}).

The $^{13}\text{CO}_2$ absorbed into the soil vapour phase was calculated as follows:

$$N_{\text{vapour}} = 5 \times (N_{\text{added}} - N_{\text{control}}) \quad (4)$$

$$N_{\text{added}} = C_{\text{CO}_2} \times \frac{V_{\text{soil air}}}{R_t \times 1000} \times \frac{(1000 + \delta^{13}\text{C}_{\text{added}}) \times R_{\text{st}}}{1000 + (1000 + \delta^{13}\text{C}_{\text{added}}) \times R_{\text{st}}} \quad (5)$$

$$N_{\text{control}} = C_{\text{CO}_2} \times \frac{V_{\text{soil air}}}{R_t \times 1000} \times \frac{(1000 + \delta^{13}\text{C}_{\text{control}}) \times R_{\text{st}}}{1000 + (1000 + \delta^{13}\text{C}_{\text{control}}) \times R_{\text{st}}} \quad (6)$$

$$V_{\text{soil air}} = \frac{V_{\text{soil}}}{1000} \times P_{\text{soil}} - \frac{V_{\text{soil}}}{1000} \times \text{VWC}, \quad (7)$$

where N_{vapour} is the amount of substance of absorbed ^{13}C in the soil vapour phase (μmol), N_{added} and N_{control} are the amounts of substance of ^{13}C in the soil vapour phase with and without $^{13}\text{CO}_2$ addition, respectively, C_{CO_2} is the concentration of CO_2 in the soil vapour phase ($\mu\text{mol mol}^{-1}$), R_t is the molar volume of gas (L mol^{-1}), $V_{\text{soil air}}$ is the volume of soil air (L), $\delta^{13}\text{C}_{\text{added}}$ and $\delta^{13}\text{C}_{\text{control}}$ are the $\delta^{13}\text{C}$ values of the soil vapour phase with and without $^{13}\text{CO}_2$ addition, respectively, R_{st} is the stable isotope ratio in the reference standard⁴⁰, V_{soil} is the volume of soil sample ($25 \text{ cm} \times 5 \text{ cm} \times 18 \text{ cm}$), P_{soil} is the soil porosity, and VWC is the mean value of soil volumetric water content during the measurement period.

The $^{13}\text{CO}_2$ absorbed into the soil liquid phase was calculated as follows:

$$N_{\text{liquid}} = \frac{5 \times (N'_{\text{added}} - N'_{\text{control}})}{4} \quad (8)$$

$$N'_{\text{added}} = \sum C_{\text{DIC}} \times \frac{V'_{\text{soil}} \times B}{1000} \times \frac{(1000 + \delta^{13}\text{C}'_{\text{added}}) \times R_{\text{st}}}{1000 + (1000 + \delta^{13}\text{C}'_{\text{added}}) \times R_{\text{st}}} \quad (9)$$

$$N'_{\text{control}} = \sum C_{\text{DIC}} \times \frac{V'_{\text{soil}} \times B}{1000} \times \frac{(1000 + \delta^{13}\text{C}'_{\text{control}}) \times R_{\text{st}}}{1000 + (1000 + \delta^{13}\text{C}'_{\text{control}}) \times R_{\text{st}}} \quad (10)$$

where N_{liquid} is the amount of substance of absorbed ^{13}C in the soil liquid phase (μmol), N'_{added} and N'_{control} are the amounts of substance of ^{13}C in the soil liquid phase with and without $^{13}\text{CO}_2$ addition, respectively, C_{DIC} is the soil bicarbonate radical concentration in the soil liquid phase ($\mu\text{mol kg}^{-1}$), V'_{soil} is the volume of soil sample ($25 \text{ cm} \times 20 \text{ cm} \times 18 \text{ cm}$), B is bulk density (g cm^{-3}), and $\delta^{13}\text{C}'_{\text{added}}$ and $\delta^{13}\text{C}'_{\text{control}}$ are the $\delta^{13}\text{C}$ values of the soil liquid phase with and without $^{13}\text{CO}_2$ addition, respectively.

A one-way analysis of variance was used to test the differences in $\delta^{13}\text{C}$, C_{DIC} , VWC and V_{leachate} between the chambers and samples. All statistical analyses were performed using MATLAB 7.12.0.635 (The Math Works, Natick, MA, USA).

References

- Delgado-Baquerizo, M. *et al.* Decoupling of soil nutrient cycles as a function of aridity in global drylands. *Nature* **502**, 672–676 (2013).
- Wiant, H. V. Has the contribution of litter decay to forest soil respiration been overestimated? *J. Forest.* **65**, 408–409 (1967).
- Raich, J. W. & Schlesinger, W. H. The global carbon dioxide flux in soil respiration and its relationship to vegetation and climate. *Tellus* **44**, 81–99 (1992).
- Jasoni, R. L., Smith, S. D. & Arnone, J. A. Net ecosystem CO_2 exchange in Mojave Desert shrublands during the eighth year of exposure to elevated CO_2 . *Global Change Bio.* **11**, 749–756 (2005).
- Wohlfahrt, G., Fenstermaker, L. F. & Arnone, J. A. Large annual net ecosystem CO_2 uptake of a Mojave Desert ecosystem. *Global Change Biol.* **14**, 1475–1487 (2008).
- Yates, E. L. *et al.* Assessing the role of alkaline soils on the carbon cycle at a playa site. *Environ. Earth Sci.* **70**, 1047–1056 (2013).
- Hamerlynck, E. P., Scott, R. L., Sánchez-Cañete, E. P. & Barron-Gafford, G. A. Nocturnal soil CO_2 uptake and its relationship to subsurface soil and ecosystem carbon fluxes in a Chihuahuan Desert shrubland. *J. Geophys. Res.* **118**, 1593–1603 (2013).
- Xie, J. X., Li, Y., Zhai, C. X., Li, C. H. & Lan, Z. D. CO_2 absorption by alkaline soil and its implication to the global carbon cycle. *Environ. Geol.* **56**, 953–961 (2009).
- Ma, J., Wang, Z. Y., Stevenson, B. A., Zheng, X. J. & Li, Y. An inorganic CO_2 diffusion and dissolution process explains negative CO_2 fluxes in saline/alkaline soil. *Sci. Rep-UK*, **3**, doi:10.1038/srep02025 (2013).
- Fa, K. Y. *et al.* CO_2 absorption of sandy soil induced by rainfall pulses in a desert ecosystem. *Hydrol. Process.* **29**, 2043–2051 (2015).
- Liu, J. B. *et al.* Abiotic CO_2 exchange between soil and atmosphere and its response to temperature. *Environ. Earth Sci.* doi: 10.1007/s12665-01403595-9 (2014).
- Ball, B. A., Virginia, R. A., Barretta, J. E., Parsons, A. N. & Wall, D. H. Interactions between physical and biotic factors influence CO_2 flux in Antarctic dry valley soils. *Soil Biol. Biochem.* **41**, 1510–1517 (2009).
- Huang, J. P., Yu, H. P., Guan, X. D., Wang, G. Y. & Guo, R. X. Accelerated dryland expansion under climate change. *Nat. Clim. Change* **6**, 166–171 (2016).
- Stone, R. Ecosystems: Have desert researchers discovered a hidden loop in the carbon cycle? *Science* **320**, 1409–1410 (2008).
- Rey, A. Mind the gap: non-biological processes contributing to soil CO_2 efflux. *Global Change Biol.* **21**, 1752–1761 (2015).
- Emmerich, W. E. Carbon dioxide fluxes in a semiarid environment with high carbonate soils. *Agr. Forest Meteorol.* **116**, 91–102 (2003).
- Cuezva, S. *et al.* Short-term CO_2 (g) exchange between a shallow karstic cavity and the external atmosphere during summer: Role of the surface soil layer. *Atmos. Environ.* **45**, 1418–1427 (2011).
- Ma, J., Liu, R., Tang, L. S., Lan, Z. D. & Li, Y. A downward CO_2 flux seems to have nowhere to go. *Biogeosciences* **11**, 10419–10450 (2014).
- Liu, J. B. *et al.* Abiotic CO_2 uptake from the atmosphere by semiarid desert soil and its partitioning into soil phases. *Geophys. Res. Lett.* **42**, doi: 10.1002/2015GL064689 (2015).
- Roland, M. *et al.* Atmospheric turbulence triggers pronounced diel pattern in karst carbonate geochemistry. *Biogeosciences* **10**, 1207–1227 (2013).
- Ryskov, Y. G., Demkin, V. A., Oleynik, S. A. & Ryskova, E. A. Dynamics of pedogenic carbonate for the last 5000 years and its role as a buffer reservoir for atmospheric carbon dioxide in soils of Russia. *Global Planet. Change* **61**, 63–69 (2008).
- Liu, Z. H. Is pedogenic carbonate an important atmospheric CO_2 sink? *Chinese Sci. Bull.* **56**, 3794–3796 (2011).
- Li, Y., Wang, Y. G., Houghton, R. A. & Tang, L. S. Hidden carbon sink beneath desert. *Geophys. Res. Lett.* **42**, doi: 10.1002/2015GL064222 (2015).
- Verrecchia, E. P. & Verrecchia, K. E. Needle-fibre calcite: a critical review and a proposed classification. *J. Sediment. Res.* **64**, 650–664 (1994).
- Verrecchia, E. P. *Pedogenic Carbonates. Encyclopedia of Geobiology-Encyclopedia of Earth Sciences Series* (Springer Science + Business Media B. V., 2011).
- Berg, A. & Banwart, S. A. Carbon dioxide mediated dissolution of Ca-feldspar: implications for silicate weathering. *Chem. Geol.* **163**, 25–42 (2000).
- Hilley, G. E., Chamberlain, C. P., Moon, S., Porder, S. & Willettd, S. D. Competition between erosion and reaction kinetics in controlling silicate-weathering rates. *Earth Planet. Sc. Lett.* **293**, 191–199 (2010).
- White, A. F. *et al.* Differential rates of feldspar weathering in granitic regoliths. *Geochim. Cosmochim. Ac.* **65**, 847–869 (2001).
- Fa, K. Y. *et al.* Patterns and possible mechanisms of soil CO_2 uptake in sandy soil. *Sci. Total Environ.* **544**, 587–594 (2016).
- Luo, Y. Q. & Zhou, X. H. *Soil Respiration and the Environment*. (Elsevier, Inc., San Diego, California, USA 2006).
- Lal, R., Kimble, J. M., Stewart, B. A. & Eswaran, H. *Global Climate Change and Pedogenic Carbonates* (CRC Press, Boca Raton: Lewis Publishers, 2000).
- Van der Hoven, S. J. & Quade, J. Tracing spatial and temporal variation in the sources of calcium in pedogenic carbonates in a semiarid environment. *Geoderma* **108**, 259–276 (2002).

33. Landi, A., Mermut, A. R. & Anderson, D. W. Origin and rate of pedogenic carbonate accumulation in Saskatchewan soils, Canada. *Geoderma* **117**, 143–156 (2003).
34. Cailleau, G., Braissant, O. & Verrecchia, E. P. Biomineralization in plants as a long-term carbon sink. *Naturwissenschaften* **91**, 191–194 (2004).
35. Feng, W. *et al.* Impact of environmental factors and biological soil crust types on soil respiration in a desert ecosystem. *PLoS One* **9**, doi: 10.1371/journal.pone.0102954 (2014).
36. Walvoord, M. A. *et al.* A reservoir of nitrate beneath desert soils. *Science* **302** 1021–1024 (2003).
37. Zhu, L. S., Gao, Y. & Ren, L. F. Isotope application in the study of relation between supply and discharge of groundwater in Yanchi of Ningxia. *Ningxia Engineering Technology* **7**, 109–115 (2008).
38. Leavitt, S. W. *et al.* Stable-carbon isotopes and soil organic carbon in wheat under CO₂ enrichment. *New Phytol.* **150**, 305–314 (2001).
39. Lu, R. K. *Methods of agrochemical soil analysis*. 90–92 (China Agricultural Science Press, Beijing, 1999).
40. Friedman, I., O’Neil, J. R. & Fleisher, M. *Compilation of stable isotope fractionation factors of geochemical interest*, in *Data of Geochemistry*, (ed. Fleischer, M.) 1–12 (U.S.G.P.O., Wash., 1977).

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Author Contributions

All authors commented on manuscript at all stages. K.Y.F. and Y.Q.Z. developed the concept of the paper and supervised the study; K.Y.F. and Z. L. performed the *in situ* measurements of the soil volumetric water content and bicarbonate radical concentration and the ¹³CO₂ tracing experiment; K.Y.F. analysed all the data and wrote the manuscript; Y.Q.Z., J.B.L. B.W., and S.G.Q. worked on each version of the manuscript and significantly contributed in its final structure and presentation.

Additional Information

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