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Biotic and Abiotic Contribution to Diurnal Soil CO_2 Fluxes from Saline/Alkaline Soils

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As the second largest carbon flux in terrestrial ecosystems, the soil CO_2 flux is closely related to the atmospheric CO_2 concentration. The soil CO_2 flux is the sum of biotic respiration and abiotic geochemical CO_2 exchange; however, little is known about abiotic CO_2 fluxes in arid areas. To investigate the relative contribution of abiotic and biotic soil CO_2 fluxes over a diurnal course, the abiotic CO_2 flux was distinguished by autoclaving sterilization in both saline and alkaline soils at an arid site in northwestern China. The results demonstrated that: (1) Over the diurnal course, the abiotic CO_2 was a significant component of the soil CO_2 flux in both saline and alkaline soil, which accounted for more than 56% of the diurnal soil CO_2 flux. (2) There was a dramatic difference in the temperature response between biotic and abiotic CO_2 fluxes: the response curves of biotic respiration were exponential in the saline soil and quadratic in the alkaline soil, while the abiotic CO_2 flux was linearly correlated with soil temperature. They were of similar magnitude but with opposite signs: resulting in almost neutral carbon emissions on daily average. (3) Due to this covering up effect of the abiotic CO_2 flux, biotic respiration was severely underestimated (directly measured soil CO_2 flux was only one-seventh of the biotic CO_2 flux in saline soil, and even an order of magnitude lower in alkaline soil). In addition, the soil CO_2 flux masked the temperature-inhibition of biotic respiration in the alkaline soil, and veiled the differences in soil biological respiration between the saline and alkaline soils. Hence, the soil CO_2 flux may not be an ideal representative of soil respiration in arid soil. Our study calls for a reappraisal of the definition of the soil CO_2 flux and its temperature dependence in arid or saline/alkaline land. Further investigations of abiotic CO_2 fluxes are needed to improve our understanding of arid land responses to global warming and to assist in identifying the underlying abiotic mechanisms.

As the second largest CO_2 flux between terrestrial ecosystems and atmosphere after photosynthesis[1], the soil CO_2 flux (F_c, the exchange rate of CO_2 from soil to atmosphere) is estimated to be 68 ± 4 Pg $C.y^{-1}$ globally[2]. A small change in F_c can significantly alter the atmospheric CO_2 concentration[2,3] and potentially amplify global warming[4–6]. Biotic soil CO_2 flux (F_b, representing soil biological respiration) ranges from 60 ± 6 $gC.m^{-2}.y^{-1}$ for tundra to 1260 ± 57 $gC.m^{-2}.y^{-1}$ for tropical moist forests[2]. Abiotic soil CO_2 flux (F_a, originating from soil abiotic processes such as carbonate weathering and CO_2 dissolution) is reported to be no more than 3–4 $gC.m^{-2}.yr^{-1}$ [7]. Thus, it is customary to assume that F_c is purely of biotic origin[8,9] and equal to the soil respiration rate[10–12]. In this context, most researchers tended to neglect F_a in F_c studies (annually or on shorter time scale)[7,13,14] and focused only on soil biotic processes[8,15,16].

However, recent studies reported anomalous fluxes or negative soil CO_2 fluxes that cannot be explained by any biotic processes of soil respiration (R_s)[17–21]. In contrast to the marginal contribution assumption, they demonstrated that F_a could significantly alter the temporal variation of F_c[22,23]. F_a might temporally dominate the terrestrial-atmosphere carbon exchange and contribute 19–68% to annual net ecosystem exchange (NEE) in semiarid shrubland[24], and even created a sink larger than 100 $gC.m^{-2}.yr^{-1}$ in desert regions[25–27]. Namely, when conditions meet, F_a can account for a significant portion of F_c: up to 13% in calcareous Mojave Desert soils[28], 40% in a Mediterranean region under dry soil conditions[18], and more than 75% in the Dry Valleys of Antarctica[29]. However, only a few studies distinguished F_a from F_c, and accurate estimates of F_a at high frequency (hourly and

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daily timescales) are more limited[22,27,29,30]. Consequently, the contribution of F_a to F_c remains poorly understood, which has resulted in an intensive debate on its magnitude and mechanisms[20,21,31–33] and thus induced biases in estimation of soil biotic processes[29,30,34]. Thus, reliable partitioning of F_a and F_b is of critical importance in quantifying their contribution to F_c. As soil biotic and abiotic processes may respond to different drivers and thus respond differently to climate change[24], this partitioning is essential in understanding the feedback of the soil carbon cycle in response to climate change[23,29,30].

Anomalous fluxes or abiotic CO_2 fluxes were mainly reported in saline or alkaline soil of arid and semiarid land[20,21,25], which may occupy 50% of the total land surfaces by the end of this century[35]. Due the imminent transition to a warmer and more arid climate[36], soils in arid and semiarid areas are generally dry and expected to become drier within this century[37–39]. Thus, studies on the biotic and abiotic components of F_c in dry land soils are needed to better understand the feedback of the carbon cycle in response to climate change in arid areas. Due to inefficient leaching resulted from low precipitation, dry land soils are usually with some degree of salinity/alkalinity. Here, we used autoclaving sterilization to distinguish F_a from F_c in saline and alkaline soils in a saline desert of northwestern China. The objectives of this study were to (1) evaluate the relative contribution of F_a and F_b to F_c over the diurnal course, and (2) quantify the bias in conventional estimation of soil biotic processes. The basic hypothesis is that in dry land saline/alkaline soils, abiotic process contributes a significant portion in CO_2 exchange between soil and atmosphere.

Materials and Methods

Site description.
Our experiments were conducted at a field site near the Fukang Station of Desert Ecology, Chinese Academy of Sciences (44°17′N, 87°56′E and 475 m a.s.l.). The station is located at the northern foot of Tianshan Mountains and the southern edge of the Gurbantunggut Desert in Northwest China, where saline and alkaline land is widely distributed[40]. The climate is temperate continental: arid, hot and dry in summer and cold in winter. Mean annual temperature is 6.6°C, mean annual precipitation is 163 mm, and mean annual class-A pan evaporation is around 800–1000 mm[41]. Soils are clay-loam in texture, with high salinity/alkalinity and low organic matter. The topography in the experiment site is flat (slope < 1°), and the groundwater table used to be very high, but has declined to a depth of 6 m in recent years. The dominant shrub is *Tamarix ramosissima* Ledeb. (average canopy cover 17%). Other herbaceous species include *Salsola nitraria* Pall., *Suaeda acuminate* Moq. and *Salicornia europaea* Linn., with canopy coverage of 5–30%, depending on the precipitation in that year.

Soil sampling.
Typically, arid land soil is highly spatially variable[42,43]. To minimize the complications resulting from high spatial variability and to attain repeatable results, we opted to use well-mixed soil samples rather than intact soil cores (repeated measurements conducted using the well-mixed soil samples were not independent, and thus a mixed effects model was used to account for the autocorrelation; see details in Data analysis and statistics section). Saline and alkaline soil (FAO/UNESCO classification: Solonchaks and Solonetz) samples (0–20 cm in depth, 12 soil cores each, i.e., a total of 24 soil cores; for each core, around 9.84 kg for alkaline soil, 6.78 kg for saline soil) were collected from a typical saline desert (around the station, bulk density 1.52 ± 0.05 g. cm^{-3}) and an alkaline site (5 km away, bulk density 1.05 ± 0.03 g.cm^{-3}), respectively. Both soils are loamy textured with low nitrate, as all the desert soils are. Given that both sampling sites contain few shallow-rooted shrubs or grass species, which rapidly decompose in this hot, arid climate, very few roots and organic debris were found in the soil samples. Each soil sample was air-dried and sieved (2-mm mesh size) to remove large stones, and kept indoor till sterilizing or measurements. The soil samples from both sites were homogenized respectively and then placed into 24 bottom-sealed stainless steel drums (21.1 cm outer diameter, 20.3 cm inner diameter and 22 cm height). For each soil type, 12 soil drums were randomly selected for subsequent autoclaving sterilization and control (6 drums per treatment). In addition, a quartz sand drum was used to test the thermal expansion and contraction effects of the soil gaseous parts.

Sterilization treatment.
To discriminate F_a from F_c, the soils were treated by autoclaving sterilization in a pressure steam chamber. Due the size of the pressure steam chamber, we sterilized one soil drum at a time. For the sterilized soil, the tops of the drums were sealed with multilayers of filter paper and brown paper to prevent water infiltrating into the soil, and then sterilization was conducted in a medical autoclave for 24 h at 120°C[44]. Then, each sterilized soil drum was placed in a UV-sterilized room to prevent microbial invasion and to allow the soil to equilibrate to the ambient temperature (room temperature) and atmospheric CO_2 before the CO_2 flux measurements started. To ensure valid comparison, the control drums filled with unsterilized soil were also covered with filter paper at the top and were maintained under ambient temperature conditions and atmospheric pressure.

CO$_2$ flux measurements.
After pre-equilibration, the soil drums were reburied in the saline field, with a 2-cm wall exposed above the soil surface to install the CO_2 flux monitoring chamber. The soil surface in the drum was at the same height as the surrounding soil to maintain its temperature in accordance with natural soil temperature fluctuation. The CO_2 flux was measured using an LI-8100 Automated Soil CO_2 Flux System (LI-COR, Lincoln, Nebraska, USA) equipped with a long-term monitoring chamber (LI-8100L). Automated measurements of CO_2 flux were made at 10-min intervals, and the time length of one measurement was set to 120 s for the low CO_2 flux rates in the arid soil. To minimize the microbial invasion effect and to maintain the sterilized soil in sterile state, the CO_2 flux measurement only lasted 1 day for all soil drums. Because we only had one LI-8100, the CO_2 fluxes for all 25 soil drums were cross measured (one drum at a time) on clear days from August 17th to October 24th 2009. Soil temperature was automatically measured at a depth of 1 cm at the same 10-min intervals using thermocouples (HTT thermocouple, OMEGA Engineering, Inc., Stamford, CT, USA), which were placed in the surrounding soil close to each drum[45]. The raw CO_2 flux data and temperature data were aggregated into hourly intervals.
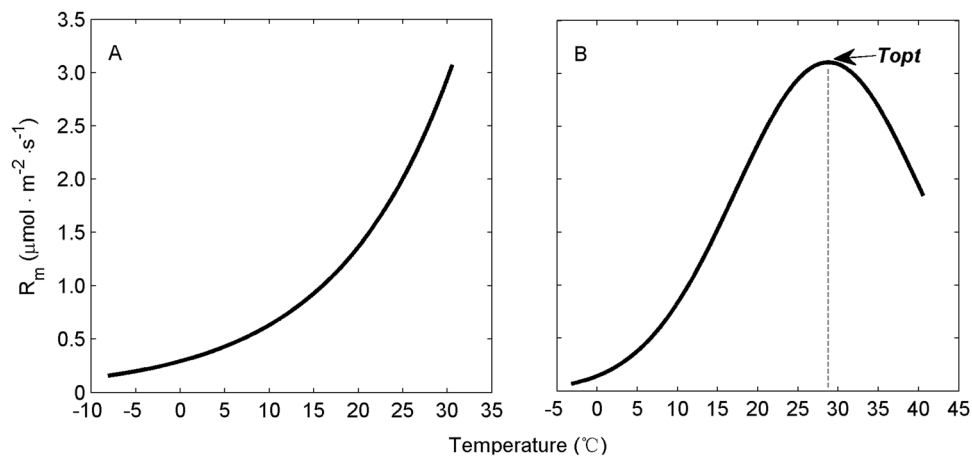
Figure 1. Schematic diagram of the relationship between soil microbial respiration (R_m) and soil temperature. (**A**) Relationship between R_m and temperature when the temperature is below the optimum temperature (T_{opt}); (**B**) Relationship between R_m and temperature when the temperature exceeds T_{opt}. The T_{opt} is defined as the temperature at which the maximum rate of soil respiration occurs.

Soil analysis.   To determine the change in the soil properties following sterilization, soil samples were collected from the soil drums after the CO_2 flux measurements were completed and analyzed for soil pH, soil electrical conductivity (EC), soil water content (SWC), soil organic carbon (SOC) and soil inorganic carbon (SIC).

Soil pH and EC were determined in a soil-water suspension (1:5 of soil: water ratio) using a potentiometer and an electric conductivity meter, respectively. SWC was determined using the conventional oven-drying and balance-weighing method. SOC was measured using the $K_2Cr_2O_7$–H_2SO_4 Walkley-Black oxidation method[46]. SIC was determined using a modified pressure transducer method[47].

Data analysis and statistics.   Based on the potential sources of CO_2, we assumed F_c is the combination of F_a and F_b:

$$F_c = F_a + F_b \tag{1}$$

F_c was determined by measuring the CO_2 flux of the control soil (unsterilized soil), representing the total CO_2 exchange rate from soil to atmosphere. F_a was determined by measuring the CO_2 flux of sterilized soil, representing the CO_2 flux resulting from soil abiotic processes. F_b was calculated as the difference between F_c and F_a.

In this study, the soils were sieved and did not contain roots; therefore, soil microbial respiration (R_m) was the main contributor to F_b. In light of previous studies[48,49], the functional relationships between R_m and temperature are demonstrated in Fig. 1. When the temperature is below the optimum temperature (T_{opt}), R_m is commonly modeled using van't Hoff equation -a simple temperature exponential function[9,10,50], or modified van't Hoff equation – using F_0 and Q_{10} as operand which was also equivalent to van't Hoff equation[50,51] (Fig. 1A). When the temperature is higher than the T_{opt}, R_m decreases with further increases in temperature due to the deactivation energy and enzyme degradation[49,52] (Fig. 1B).

$$\text{van't Hoff: } F_b = R_m = \alpha e^{\beta T} \left(T \leq T_{opt} \right) \tag{2}$$

$$\text{Modified van't Hoff: } F_b = F_0 \, Q_{10}^{\frac{(T - T_0)}{10}} \tag{3}$$

$$Q_{10} = e^{\beta \times 10} \tag{4}$$

where α and β are coefficients estimated by non-linear regression: α denotes the reference soil respiration at 0°C and β provides an estimate of the Q_{10} coefficient (Eq. 4), representing the degree of dependence of soil respiration on temperature; T is soil temperature (°C); and T_{opt} is the optimum soil temperature(°C); where Q_{10} is temperature sensitivity, defined as the factor by which CO_2 production increases with a 10°C rise in temperature; F_0 is a basal respiration rate for the temperature T_0.

Therefore, the functional relationship between F_b and soil temperature was used to validate the differentiation ability of the sterilization method. Diurnal patterns of F_c, F_a and F_b in the saline and alkaline soils were constructed by using the mean hourly CO_2 flux measured in 6 drums. The contribution of F_a to F_c was quantified as the ratio between the absolute value of F_a and the sum of the absolute values of F_a and F_b determined as mean hourly values. The hourly CO_2 flux and temperature data were used to establish the relationships between soil temperature and F_c and F_a, while the mean hourly data from 6 drums were used in the relationship of F_b to soil temperature.

In the current study, repeated measurements at the same location over time are not independent and thus need to be corrected for autocorrelation. This can be done using mixed effects modeling[53]. Therefore, a mixed

Soil type and treatment	pH (1:5)	EC (1:5) (mS/cm)	SWC (%)	SOC (%)	SIC (%)
Control saline soil	8.86 ± 0.04^a	10.33 ± 0.04^a	3.36 ± 0.05^a	1.32 ± 0.04^a	1.08 ± 0.02^a
Sterilized saline soil	8.85 ± 0.05^a	10.37 ± 0.05^a	3.34 ± 0.06^a	1.31 ± 0.02^a	1.10 ± 0.01^a
Control alkaline soil	10.32 ± 0.01^b	1.96 ± 0.02^b	0.94 ± 0.04^b	0.26 ± 0.02^b	0.94 ± 0.01^b
Sterilized alkaline soil	10.31 ± 0.02^b	1.95 ± 0.01^b	0.96 ± 0.05^b	0.27 ± 0.02^b	0.94 ± 0.01^b

Table 1. Soil properties of the control and sterilized saline and alkaline soils. Values represent the mean $\pm$ SE (n = 6). Different letters indicate significant differences (p < 0.05) between soil type and treatments based on Student's t-tests.

effects model was used to compare the diurnal patterns of F_c, F_a and F_b. Linear and non-linear regression analyses were used to statistically quantify the relationships between CO_2 flux and soil temperature. Significance level was set at the 5%. All statistical analyses were performed in SAS version 9.1 using Proc Mixed (SAS Institute Inc., 2004). The figures were drawn using the MATLAB, R2012a mapping software (The MathWorks Inc., USA.).

Results

Conservation of soil properties during sterilization. The properties of the control and sterilized saline and alkaline soils are presented in Table 1. There were no significant differences in the soil properties between the control and sterilized soils for each soil type (p > 0.05), which means that the sterilizing process did not alter the soil properties. Compared with the alkaline soil, before and after the sterilization, the saline soil had higher EC, SWC, SOC, SIC and lower pH values (Table 1, p < 0.05).

Diurnal patterns of F_c, F_a and F_b. The mean hourly CO_2 fluxes (and their standard errors) of the control and sterilized soils revealed stable diurnal variations of F_c and F_a (Fig. 2). F_c showed a pronounced unimodal diurnal pattern in both the saline and alkaline soils; the maximum value occurred at 10:00–11:00 h and the minimum at 2:00 h. While correlated with soil temperature, the diurnal pattern of F_c preceded that of soil temperature by 3 h. In the saline soil, the diurnal amplitude of F_c was 3.4 μmol.m^{-2}.s^{-1} (−0.97 to 2.43 μmol.m^{-2}.s^{-1}), whereas in the alkaline soil, the diurnal amplitude of F_c was 1.58 μmol.m^{-2}.s^{-1} (−0.51 to 1.07 μmol.m^{-2}.s^{-1}). Over the diurnal course, F_c was positive (CO_2 released to the atmosphere) from 8:00–17:00 h but negative (CO_2 taken up from the atmosphere) for the rest of the day in both the saline and alkaline soils. Despite the diurnal variation in temperature, the F_c measured in the quartz sand fluctuated around zero throughout the day (Fig. 2C).

Similar to F_c, the unimodal diurnal pattern of F_a varied from the minimum value at 22:00–0:00 h to a maximum at 11:00–12:00 h with smaller amplitude (Fig. 2). However, the peak value of F_a occurred 2 h earlier compared with soil temperature in both the saline and alkaline soils. The diurnal amplitude of F_a was 2.18 μmol.m^{-2}.s^{-1} (−1.34 to 0.84 μmol.m^{-2}.s^{-1}) in the saline soil and 1.42 μmol.m^{-2}.s^{-1} (−0.82 to 0.60 μmol.m^{-2}.s^{-1}) in the alkaline soil. F_a was consistently negative over the diurnal course except from 10:00–16:00 h in the saline soil and 9:00–15:00 h in the alkaline soil. On hourly scale, the average contribution of F_a to F_c was 56% (range of 11–79%) in the saline soil and 58% (range of 13–77%) in the alkaline soil, indicating that F_a rivaled or even exceeded F_b in contributing to F_c over the diurnal course.

F_b was obtained by subtracting F_a from F_c. The calculated F_b was consistently positive over the diurnal course in both the saline and alkaline soils (Fig. 3). In the saline soil, F_b exhibited a unimodal diurnal pattern, which preceded the soil temperature by 4 h. The maximum value of F_b in the saline soil was 1.94 μmol.m^{-2}.s^{-1} at 10:00 h and the minimum was 0.35 μmol.m^{-2}.s^{-1} at 1:00 h. However, the diurnal pattern of F_b in the alkaline soil was significantly different from that in the saline soil. In the alkaline soil, the diurnal variation of F_b indicated a bi-modal pattern. In the alkaline soil, F_b increased in the morning with increasing soil temperature and reached the first peak at 10:00 h (0.68 μmol.m^{-2}.s^{-1}). However, as the soil temperature continued to increase, F_b exhibited a small dip from 11:00–16:00 h, and then a second peak appeared at 17:00 h (0.63 μmol.m^{-2}.s^{-1}). F_b decreased throughout the night as the soil temperature decreased and reached a minimum value of 0.14 μmol.m^{-2}.s^{-1} at 7:00 h. The diurnal amplitude of F_b was 1.59 μmol.m^{-2}.s^{-1} in the saline soil and 0.54 μmol.m^{-2}.s^{-1} in the alkaline soil.

The daily averages of F_c, F_a and F_b were compared to estimate the contribution of F_a and F_b to F_c on a daily scale (Fig. 4). The daily average of F_c was 0.10 μmol.m^{-2}.s^{-1} in the saline soil, which was only slightly higher than the 0.00 μmol.m^{-2}.s^{-1} value for the alkaline soil. The average daily F_a in the saline soil was −0.63 μmol.m^{-2}.s^{-1}, which was approximately 170% greater than the value measured in the alkaline soil (−0.37 μmol.m^{-2}.s^{-1}). Similarly, the average daily F_b in the saline soil was 195% greater than the value measured in the alkaline soil (0.72 and 0.37 μmol.m^{-2}.s^{-1}, respectively). The average daily F_a and F_b values were of a similar magnitude but with different signs, indicating that the CO_2 released by F_b was largely offset by the CO_2 taken up by F_a over a diurnal course. Moreover, due to this covering up effect of F_a, F_c was only one-seventh of F_b in saline soil, and even an order of magnitude lower than F_b in alkaline soil, which resulted in almost neutral carbon emissions over the course of a day.

Temperature control of F_c, F_a and F_b. The correlations between F_c and soil temperature were significant for both the saline (Fig. 5A, p < 0.01) and alkaline soils (Fig. 5B, p < 0.01), and soil temperature explained 81% and 67% of the variation in F_c in the saline and alkaline soils, respectively. Similarly, soil temperature explained 88% and 65% of the variation in F_a in the saline and alkaline soils, respectively. Furthermore, according to the
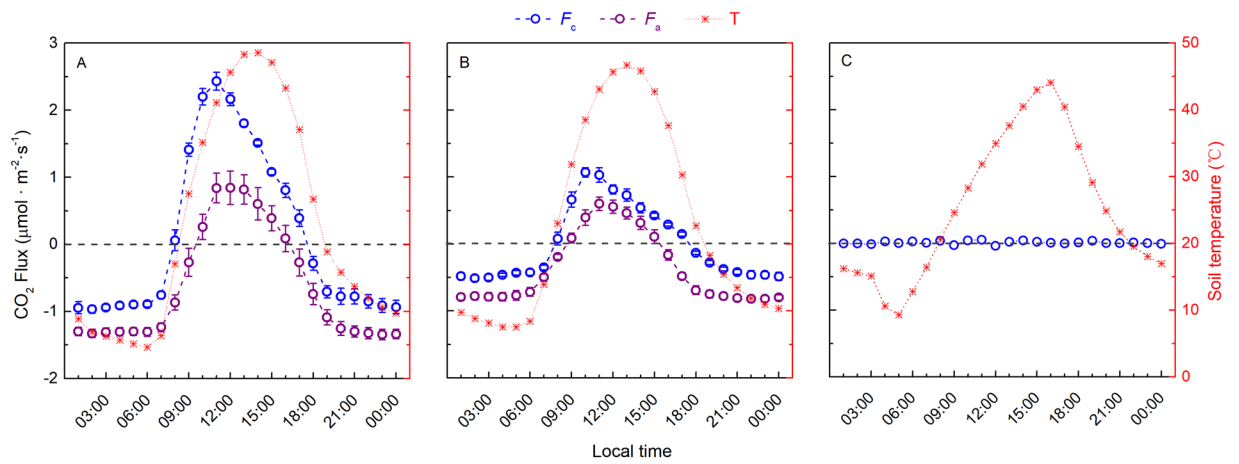
Figure 2. The diurnal patterns of soil CO_2 flux (F_c), abiotic CO_2 flux (F_a) and soil temperature. (**A**) Saline soil; (**B**) alkaline soil; (**C**) quartz sand. Soil temperature was measured at a depth of 1 cm. Each value represents the mean $\pm$ SE, n = 6, except for quartz sand (n = 1).
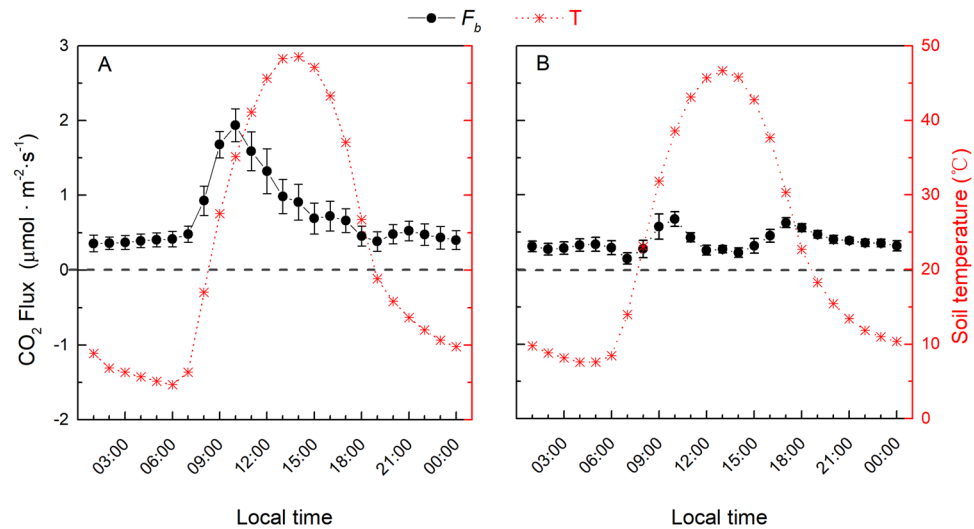

Figure 3. The diurnal patterns of biotic CO_2 flux (F_b) and soil temperature. (**A**) Saline soil; (**B**) alkaline soil. Soil temperature was measured at a depth of 1 cm. Each value represents the mean $\pm$ SE, n = 6.
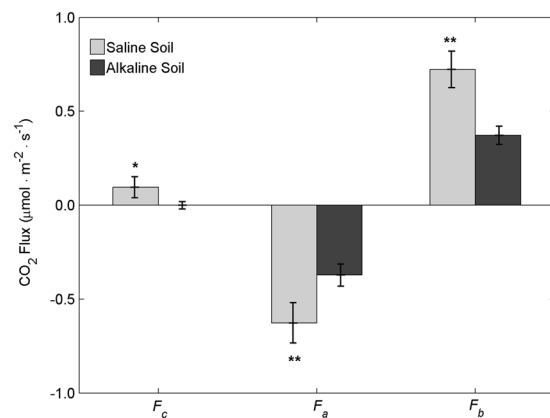
Figure 4. The daily averages of F_c, F_a and F_b in the saline and alkaline soils. Each value represents the mean $\pm$ SE, n = 6. [*]and [**]denote significant differences between the soils at $\alpha = 0.05$ and $\alpha = 0.01$, respectively.
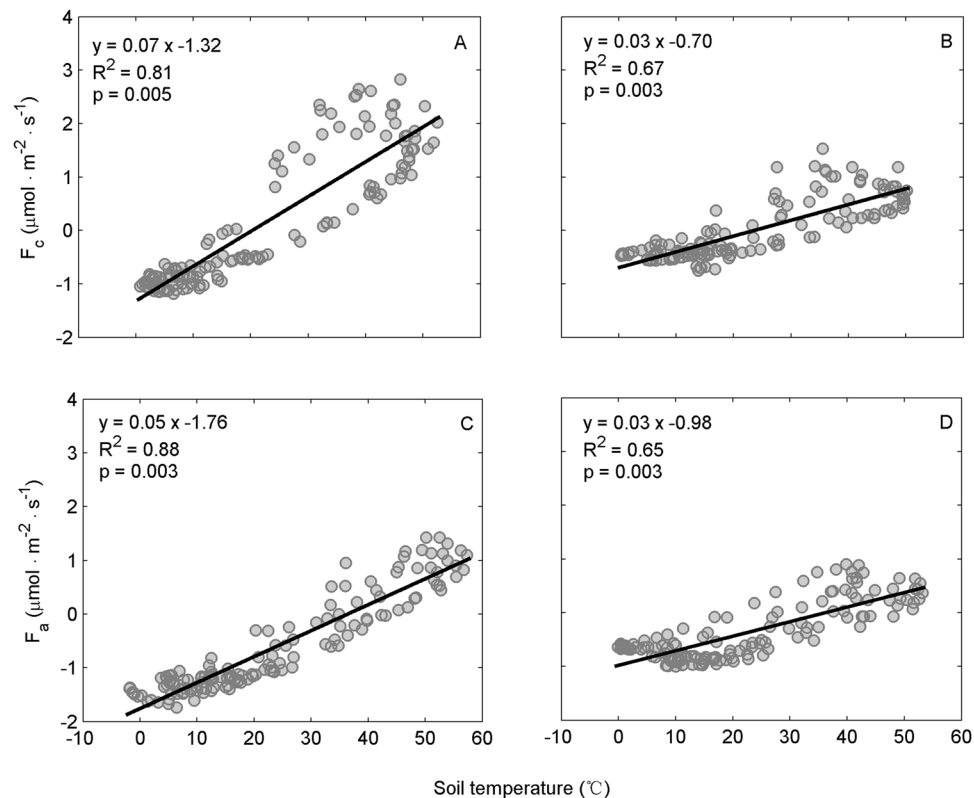
Figure 5. Relationships between soil temperature and hourly F_c and F_a. (**A,C**) Salinesoil; (**B,D**) alkaline soil. Soil temperature was measured at a depth of 1 cm.
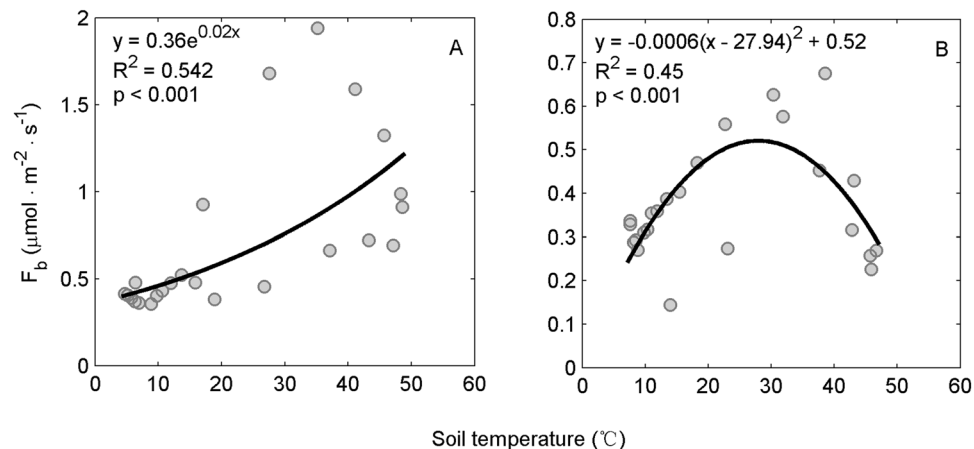

Figure 6. Relationship between soil temperature and F_b. (**A**) Saline soil; (**B**) alkaline soil. Soil temperature was measured at a depth of 1 cm.

fitted F_a–temperature function, there was a threshold temperature of approximately 36 °C, for both the saline and alkaline soils, where F_a changed from negative to positive with an increase in soil temperature.

The relationship between hourly F_b and soil temperature in the saline soil was significantly different from that in the alkaline soil (Fig. 6). In the saline soil, F_b increased with an increase in soil temperature (Fig. 6A). An exponential function yielded the best fit for the relationship between F_b and soil temperature, explaining 54% of the variation in F_b in the saline soil over the diurnal course. However, in the alkaline soil, F_b increased with an increase in soil temperature until 28 °C, followed by a decrease with further increases in soil temperature (Fig. 6B). Thus, the effect of soil temperature on F_b in the alkaline soil was best described using a quadratic function, and soil temperature explained 45% of the diurnal variation in F_b. In general, the value of F_b was lower in the alkaline soil than in the saline soil at the same soil temperature.

Discussion

The contribution of abiotic CO₂ flux. The soil CO_2 fluxes reported in the current study (Figs. 2 and 4) are comparable in magnitude to soil CO_2 flux measured *in situ* in the same undisturbed saline desert region[27] and in alkaline soil[54], suggesting that soil sampling procedures have little effect on soil CO_2 fluxes. Previous studies showed that air-drying and sieving of soil could significantly affect soil CO_2 fluxes by changing the soil water content, soil structure and soil organic matter fractions[55,56]. The arid soil used in the current study was severely dry, low in organic matter content and loose in structure, and our results are similar to *in situ* soil measurements[57]. We obtained continuous and high frequency measurements of abiotic CO_2 fluxes covering diurnal courses after autoclaving sterilization (Fig. 2) without altering soil chemical properties (Table 1), although we have to admit that steaming may change soil moisture of the sterilized soil, but in our case, this change is not significant (Table 1). In addition, soil abiotic reactions and its rate depended on the concentration of CO_2 in the soil air, and after sterilization, soil biological activity was eliminated and thus stopped CO_2 release into the soil. In this case, the change of soil CO_2 concentration might lead to the overestimation or underestimation of the abiotic CO_2 flux (shifted towards values that are more or less positive). In original soils, positive respiration means its CO_2 concentration higher than atmosphere, while negative respiration means its CO_2 lower than atmosphere. Therefore, autoclaving sterilization is not an accurate way to distinguish abiotic CO_2 flux from soil CO_2 flux, but only the best way we can find after trying various method of sterilizing the soil. The results obtained can be considered as reference values not far from the truth.

Contrary to the marginal contribution over short timescales in a previous estimation[7], our results demonstrated that abiotic CO_2 flux accounted for more than 56% of the soil CO_2 flux measured over diurnal courses in both the saline and alkaline soils (Figs. 2 and 4). The results also show that soil temperature exerted a dominant control over abiotic CO_2 flux (Fig. 5), which is consistent with recent studies in Antarctica[22,29]. Thus, as an important component of soil CO_2 flux, abiotic CO_2 flux should be considered in arid ecosystem carbon budgets, which is predicted to be more susceptible to climate change[36,58].

The current study also helps to clarify the controversy regarding the underlying mechanism of the abiotic CO_2 flux[20,21,32,33]. First, photosynthesis in the autotrophic community in biological soil crusts was ruled out because sterilization eliminated soil biological activity. According to recent studies, the main abiotic interpretations involved were subterranean ventilation[20,23,59], carbonate weathering[17–19,60], and CO_2 dissolution in soil water[22,44,61]. The zero CO_2 flux measured in the quartz sand and the comparable soil CO_2 flux measured *in situ* in the saline and alkaline soils (Fig. 2) illustrated that in the current study, subterranean ventilation resulting from thermal expansion and contraction of the soil gasphase could not be the main contributor of the abiotic CO_2 flux. Moreover, the abiotic CO_2 flux was positively correlated with temperature (bidirectional - CO_2 was released at higher temperature and absorbed at lower temperature), and there was a difference in this process between the saline and alkaline soils (Fig. 5). Therefore, the underlying abiotic processes were temperature-regulated, reversible, physical-chemical processes that were also affected by soil salinity and alkalinity. Although both carbonate weathering and CO_2 dissolution in soil water conform to these characteristics, the CO_2 flux resulting from carbonate weathering is relatively small[20,34,62] compared with our data. In contrast, the DIC (dissolved inorganic carbon) derived from soil CO_2 dissolution is quite large and comparable to daily soil CO_2 flux[34,40]. The modeled CO_2 dissolution process produced CO_2 flux values that were comparable to our results[44]. Thus, CO_2 dissolution was the most likely mechanism underlying the abiotic CO_2 flux. Although the current study used air-dried soils with a limited amount of water, but the dryness accelerated the salinity and alkalinity of the soils, which by nature, possesses a high CO_2 dissolution ability[44]. On the other hand, we admit that further studies are needed to draw a concrete conclusion on the underlying abiotic mechanisms.

The difference between soil respiration and measured soil CO₂ flux. As the soil abiotic processes produced considerable CO_2 that was absorbed by/emitted from the soil during the diurnal courses (Fig. 3), direct measurements of the soil CO_2 flux can be considerably different from the biological respiration rate in saline/alkaline soils. The biotic CO_2 flux we obtained was positive throughout the day (Fig. 3), which confirms the nature of biological respiration (unidirectional release of CO_2 to the atmosphere)[8,9]. Moreover, the higher biotic CO_2 flux in the saline than in the alkaline soil (Fig. 4) reflected its relatively higher soil water content and soil organic carbon content (Table 1). The biotic CO_2 flux–temperature relationships (Fig. 6) agreed well with previous reports[49,63,64]. The 28 °C optimum temperature for soil respiration was also comparable to synthesis studies conducted in seven deserts[65].

Our results showed that, on daily average, the soil CO_2 flux was only one-seventh of the biotic CO_2 flux in saline soil, and even an order of magnitude lower in alkaline soil, due to the negative value of the abiotic CO_2 flux (Fig. 4).Hence, the soil CO_2 flux severely underestimated the biotic CO_2 contribution. This result agreed well with a study in a semiarid soil where biological respiration was 3.8 times higher than the measured soil CO_2 flux[30,34] but was different to a study in Antarctic dry valley soils where the soil CO_2 flux measurements overestimated the biotic CO_2 flux[29]. In addition, the temperature response curves of the soil CO_2 flux and the biotic CO_2 flux were dramatically different. While the soil CO_2 fluxes were linearly correlated with soil temperature (Fig. 5), the biotic CO_2 flux–temperature relationships were exponential in the saline soil (Fig. 6A) and quadratic in the alkaline soil (Fig. 6B). Accordingly, using soil CO_2 flux to represent the biotic respiration of arid soil would misestimate the temperature response of soil biota and mask the temperature-inhibition effect in alkaline soil, which ultimately leads to considerable bias in the responses of arid soil to climate change. The temperature-inhibition of biotic respiration in alkaline soil might come from the effect of drought[51]. Therefore, it is noteworthy that in other ecosystems where environmental conditions are more favorable (e.g., higher soil water content, higher soil organic content, more roots), the relative contribution of the abiotic CO_2 flux would decrease with considerable increase in the soil biology activity. Moreover, the average daily soil CO_2 flux in the saline soil was only slightly higher than in the alkaline soil, while the biotic CO_2 flux in the saline soil was approximately 2-times higher than in the

alkaline soil(Fig. 4), indicating the soil CO_2 flux measurements also underestimated soil biological activity differences between the saline and alkaline soils.

In general, the soil CO_2 flux measured in arid soil is not an ideal reflection of soil respiration over the diurnal course. In addition, it is important to distinguish the abiotic or biotic sources of soil CO_2 flux, so that estimates of the temperature response of soil biotic CO_2 flux and the dynamics of soil organic matter in arid soils are not obscured or underestimated by soil CO_2 flux measurements due to co-varying soil abiotic processes.

Conclusions

The current study is an attempt to distinguish abiotic contribution from the soil CO_2 flux, by autoclaving sterilization and quantifying the abiotic contributions over diurnal courses in saline and alkaline soils. The results demonstrated that the abiotic flux was an important component of the soil CO_2 flux in both the saline and alkaline soils. If taken the directly measured soil CO_2 flux as soil respiration, soil biological respiration might be underestimated in the saline and alkaline soils. Moreover, the dramatic difference in the temperature response between biotic and abiotic CO_2 fluxes suggested that the responses of the soil CO_2 flux in arid land are not simple, but a combined results of co-varied soil biotic and abiotic processes. Our study calls for a reappraisal of the understanding of the soil CO_2 flux and its temperature dependence in arid or saline/alkaline land.

Data availability

All relevant data is contained within the manuscript.

References

1. Schimel, D. S. Terrestrial ecosystems and the carbon-cycle. *Glob. Change Biol.* **1**, 77–91, https://doi.org/10.1111/j.1365-2486.1995.tb00008.x. (1995).
2. Raich, J. W. & Schlesinger, W. H. The global carbon dioxide flux in soil respiration and its relationship to vegetation and climate. *Tellus B* **44**, 81–99, https://doi.org/10.1034/j.1600-0889.1992.t01-1-00001.x (1992).
3. Schlesinger, W. H. & Andrews, J. A. Soil respiration and the global carbon cycle. *Biogeochemistry* **48**, 7–20, https://doi.org/10.1023/A:1006247623877 (2000).
4. Cox, P. M., Betts, R. A., Jones, C. D., Spall, S. A. & Totterdell, I. J. Acceleration of global warming due to carbon-cycle feedbacks in a coupled climate model. *Nature* **408**, 184–187, https://doi.org/10.1038/35041539 (2000).
5. Giardina, C. P. & Ryan, M. G. Evidence that decomposition rates of organic carbon in mineral soil do not vary with temperature. *Nature* **404**, 858–861, https://doi.org/10.1038/35009076 (2000).
6. Luo, Y., Wan, S., Hui, D. & Wallace, L. L. Acclimatization of soil respiration to warming in a tall grass prairie. *Nature* **413**, 622–625, https://doi.org/10.1038/35098065 (2001).
7. Kuzyakov, Y. Sources of CO_2 efflux from soil and review of partitioning methods. *Soil Biol. Biochem.* **38**, 425–448, https://doi.org/10.1016/j.soilbio.2005.08.020 (2006).
8. Bond-Lamberty, B. & Thomson, A. Temperature-associated increases in the global soil respiration record. *Nature* **464**, 579–582, https://doi.org/10.1038/nature08930 (2010).
9. Lloyd, J & Taylor, J. A. On the temperature dependence of soil respiration. *Funct. Ecol* **8**, 315–323, https://doi.org/10.2307/2389824 (1994).
10. Fang, C. & Moncrieff, J. B. A model for soil CO_2 production and transport 1: Model development. *Agric. For. Meteorol* **95**, 225–236, https://doi.org/10.1016/S0168-1923(99)00036-2 (1999).
11. Jensen, L. S. *et al.* Soil surface CO_2 flux as an index of soil respiration *in situ*: A comparison of two chamber methods. *Soil Biol. Biochem.* **28**, 1297–1306, https://doi.org/10.1016/S0038-0717(96)00136-8 (1996).
12. Selsted, M. B. *et al.* Soil respiration is stimulated by elevated CO_2 and reduced by summer drought: three years of measurements in a multifactor ecosystem manipulation experiment in a temperate heathland (CLIMAITE). *Glob. Change Biol.* **18**, 1216–1230, https://doi.org/10.1111/j.1365-2486.2011.02634.x (2012).
13. Schlesinger, W. H. The formation of caliche in soils of the Mojave Desert, California. *Geochim. Cosmochim. Acta* **49**, 57–66, https://doi.org/10.1016/0016-7037(85)90191-7 (1985).
14. Schlesinger, W. H. *Biogeochemistry: an analysis of global change.* (Academic press, 1997).
15. Wang, Y. G., Hong, Z. & Yan, L. Spatial heterogeneity of soil moisture, microbial biomass carbon and soil respiration at stand scale of an arid scrubland. *Environmental Earth Sciences* **70**, 3217–3224, https://doi.org/10.1007/s12665-013-2386-z (2013).
16. Li, Y., Hou, C., Song, C. & Guo, Y. Seasonal changes in the contribution of root respiration to total soil respiration in a freshwater marsh in Sanjiang Plain, Northeast China. *Environmental Earth Sciences* **75**, 848, https://doi.org/10.1007/s12665-016-5592-7 (2016).
17. Emmerich, W. E. Carbon dioxide fluxes in a semiarid environment with high carbonate soils. *Agric. For. Meteorol* **116**, 91–102, https://doi.org/10.1016/S0168-1923(02)00231-9 (2003).
18. Inglima, I. *et al.* Precipitation pulses enhance respiration of Mediterranean ecosystems: the balance between organic and inorganic components of increased soil CO_2 efflux. *Glob. Change Biol.* **15**, 1289–1301, https://doi.org/10.1111/j.1365-2486.2008.01793.x (2009).
19. Mielnick, P., Dugas, W. A., Mitchell, K. & Havstad, K. Long-term measurements of CO_2 flux and evapotranspiration in a Chihuahuan desert grassland. *Journal of Arid Environments* **60**, 423–436, https://doi.org/10.1016/j.jaridenv.2004.06.001 (2005).
20. Serrano-Ortiz, P. *et al.* Hidden, abiotic CO_2 flows and gaseous reservoirs in the terrestrial carbon cycle: Review and perspectives. *Agric. For. Meteorol* **150**, 321–329, https://doi.org/10.1016/j.agrformet.2010.01.002 (2010).
21. Stone, R. Have desert researchers discovered a hidden loop in the carbon cycle? *Science* **320**, 1409–1410, https://doi.org/10.1126/science.320.5882.1409 (2008).
22. Ball, B. A., Virginia, R. A., Barrett, 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, https://doi.org/10.1016/j.soilbio.2009.04.011 (2009).
23. Kowalski, A. S. *et al.* Can flux tower research neglect geochemical CO_2 exchange? *Agric. For. Meteorol* **148**, 1045–1054, https://doi.org/10.1016/j.agrformet.2008.02.004 (2008).
24. Serrano-Ortiz, P. *et al.* Interannual CO_2 exchange of a sparse Mediterranean shrubland on a carbonaceous substrate. *Journal of Geophysical Research: Biogeosciences* **114**, https://doi.org/10.1029/2009JG000983 (2009).
25. 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. *Glob. Change Biol.* **11**, 749–756, https://doi.org/10.1111/j.1365-2486.2005.00948.x (2005).
26. Wohlfahrt, G., Fensternaker, L. F. & Arnone, J. A. Large annual net ecosystem CO_2 uptake of a Mojave Desert ecosystem. *Glob. Change Biol.* **14**, 1475–1487, https://doi.org/10.1111/j.1365-2486.2008.01593.x (2008).

27. Xie, J., Li, Y., Zhai, C., Li, C. & Lan, Z. CO₂ absorption by alkaline soils and its implication to the global carbon cycle. *Environmental Geology* **56**, 953–961, https://doi.org/10.1007/s00254-008-1197-0 (2009).

28. Stevenson, B. A. & Verburg, P. S. J. Effluxed CO₂-¹³C from sterilized and unsterilized treatments of a calcareous soil. *Soil Biol. Biochem.* **38**, 1727–1733, https://doi.org/10.1016/j.soilbio.2005.11.028 (2006).

29. Shanhun, F. L., Almond, P. C., Clough, T. J. & Smith, C. M. S. Abiotic processes dominate CO₂ fluxes in Antarctic soils. *Soil Biol. Biochem.* **53**, 99–111, https://doi.org/10.1016/j.soilbio.2012.04.027 (2012).

30. Fa, K. *et al.* Underestimation of soil respiration in a desert ecosystem. *CATENA* **162**, 23–28, https://doi.org/10.1016/j.catena.2017.11.019 (2018).

31. Schlesinger, W. H. An evaluation of abiotic carbon sinks in deserts. *Glob. Change Biol.* **23**, 25–27, https://doi.org/10.1111/gcb.13336 (2017).

32. Schlesinger, W. H., Belnap, J. & Marion, G. On carbon sequestration in desert ecosystems. *Glob. Change Biol.* **15**, 1488–1490, https://doi.org/10.1111/j.1365-2486.2008.01763.x (2009).

33. Fa, K. Y. *et al.* Patterns and possible mechanisms of soil CO₂ uptake in sandy soil. *Science of the Total Environment* **544**, 587–594, https://doi.org/10.1016/j.scitotenv.2015.11.163 (2016).

34. Angert, A. *et al.* Using O₂ to study the relationships between soil CO₂ efflux and soil respiration. *Biogeosciences Discussions* **11**, 12039–12068, https://doi.org/10.5194/bg-12-2089-2015 (2014).

35. Huang, J., Yu, H., Guan, X., Wang, G. & Guo, R. Accelerated dryland expansion under climate change. *Nature Climate Change* **6**, 166–171, https://doi.org/10.1038/nclimate2837 (2015).

36. Chapin, F. S., Randerson, J. T., McGuire, A. D., Foley, J. A. & Field, C. B. Changing feedbacks in the climate-biosphere system. *Front. Ecol. Environ.* **6**, 313–320, https://doi.org/10.1890/080005 (2008).

37. Biasutti, M. Climate change: Future rise in rain inequality. *Nat. Geosci.* **6**, 337–338, https://doi.org/10.1038/ngeo1814 (2013).

38. Borken, W. & Matzner, E. Reappraisal of drying and wetting effects on C and N mineralization and fluxes in soils. *Glob. Change Biol.* **15**, 808–824, https://doi.org/10.1111/j.1365-2486.2008.01681.x (2009).

39. Huang, J., Yu, H., Dai, A., Wei, Y. & Kang, L. Drylands face potential threat under 2 °C global warming target. *Nature Climate Change* https://doi.org/10.1038/nclimate3275 (2017).

40. Wang, Y., Wang, Z. & Li, Y. Storage/turnover rate of inorganic carbon and its dissolvable part in the profile of saline/alkaline soils. *PLoS One* **8**, e82029, https://doi.org/10.1371/journal.pone.0082029 (2013).

41. Xu, G.-Q., Yu, D.-D. & Li, Y. Patterns of biomass allocation in Haloxylon persicum woodlands and their understory herbaceous layer along a groundwater depth gradient. *Forest Ecology and Management* **395**, 37–47 (2017).

42. Maestre, F. T. & Cortina, J. Small-scale spatial variation in soil CO₂ efflux in a Mediterranean semiarid steppe. *Appl. Soil Ecol* **23**, 199–209, https://doi.org/10.1016/S0929-1393(03)00050-7 (2003).

43. Pen-Mouratov, S., Rakhimbaev, M. & Steinberger, Y. Spatio-temporal effect on soil respiration in fine-scale patches in a desert ecosystem. *Pedosphere* **16**, 1–9, https://doi.org/10.1016/S1002-0160(06)60019-2 (2006).

44. Ma, J., Wang, Z. Y., Stevenson, B. A., Zheng, X. J. & Li, Y. An inorganic CO₂ diffusion and dissolution process explains negative CO₂ fluxes in saline/alkaline soils. *Sci Rep* **3**, 2025, https://doi.org/10.1038/srep02025 (2013).

45. Ruehr, N. K., Knohl, A. & Buchmann, N. Environmental variables controlling soil respiration on diurnal, seasonal and annual time-scales in a mixed mountain forest in Switzerland. *Biogeochemistry* **98**, 153–170, https://doi.org/10.1007/s10533-009-9383-z (2010).

46. Nelson, D. W. & Sommers, L. E. Total carbon, organic carbon, and organic matter. *Methods of soil analysis. Part 3-chemical methods.* **3**, 961–1010, https://doi.org/10.2136/sssabookser5.3.c34 (1996).

47. Sherrod, L. A., Dunn, G., Peterson, G. A. & Kolberg, R. L. Inorganic carbon analysis by modified pressure-calcimeter method. *Soil Sci. Soc. Am. J* **66**, 299–305, https://doi.org/10.2136/sssaj2002.2990 (2002).

48. Luo, Y. & Zhou, X. *Soil respiration and the environment.* (San Diego, CA, USA: Academic Press/Elsevier., 2006).

49. Richardson, J., Chatterjee, A. & Jenerette, G. D. Optimum temperatures for soil respiration along a semi-arid elevation gradient in southern California. *Soil Biol. Biochem.* **46**, 89–95, https://doi.org/10.1016/j.soilbio.2011.11.008 (2012).

50. Davidson, E. A., Janssens, I. A. & Luo, Y. Q. On the variability of respiration in terrestrial ecosystems: moving beyond Q₁₀. *Glob. Change Biol.* **12**, 154–164 (2006).

51. Davidson, E. A. & Janssens, I. A. Temperature sensitivity of soil carbon decomposition and feedbacks to climate change. *Nature* **440**, 165–173, https://doi.org/10.1038/Nature04514 (2006).

52. Flanagan, P. & Veum, A. Relationships between respiration, weight loss, temperature and moisture in organic residues on tundra. *Soil organisms and decomposition in tundra*, 249–277 (1974).

53. Xie, J.-B., Xu, G.-Q., Cao, X., Wang, Z.-Y. & Li, Y. When the classical reaction norm is corrected by body size. *Perspectives in Plant Ecology Evolution & Systematics* **17**, 454–466, https://doi.org/10.1016/j.ppees.2015.09.007 (2015).

54. Li, Y., Wang, Y.-G., Houghton, R. A. & Tang, L.-S. Hidden carbon sink beneath desert. *Geophysical Research Letters* **42**, 5880–5887, https://doi.org/10.1002/2015GL064222 (2015).

55. Datta, R., Vranová, V., Pavelka, M., Rejšek, K. & Formánek, P. Effect of soil sieving on respiration induced by low-molecular-weight substrates. *Int. Agrophys.* **28**, 119–124, https://doi.org/10.2478/intag-2013-0034 (2014).

56. Six, J., Conant, R. T., Paul, E. A. & Paustian, K. Stabilization mechanisms of soil organic matter: Implications for C-saturation of soils. *Plant and soil* **241**, 155–176, https://doi.org/10.1023/a:1016125726789 (2002).

57. Xu, H. & Li, Y. Water-use strategy of three central Asian desert shrubs and their responses to rain pulse events. *Plant and soil* **285**, 5–17, https://doi.org/10.1007/s11104-005-5108-9 (2006).

58. Ji, F., Wu, Z., Huang, J. & Chassignet, E. P. Evolution of land surface air temperature trend. *Nature Climate Change* **4**, 462–466, https://doi.org/10.1038/nclimate2223 (2014).

59. Were, A. *et al.* Ventilation of subterranean CO₂ and Eddy covariance incongruities over carbonate ecosystems. *Biogeosciences* **7**, 859–867, https://doi.org/10.5194/bg-7-859-2010 (2010).

60. Soper, F. M., McCalley, C. K., Sparks, K. & Sparks, J. P. Soil carbon dioxide emissions from the Mojave desert: Isotopic evidence for a carbonate source. *Geophysical Research Letters* **44**, 245–251, https://doi.org/10.1002/2016gl071198 (2017).

61. Parsons, A. N., Barrett, J. E., Wall, D. H. & Virginia, R. A. Soil carbon dioxide flux in Antarctic dry valley ecosystems. *Ecosystems* **7**, 286–295, https://doi.org/10.1007/s10021-003-0132-1 (2004).

62. Roland, M. *et al.* Atmospheric turbulence triggers pronounced diel pattern in karst carbonate geochemistry. *Biogeosciences Discussions* **10**, 1207–1227, https://doi.org/10.5194/bg-10-5009-2013 (2013).

63. Davidson, E. A., Belk, E. & Boone, R. D. Soil water content and temperature as independent or confounded factors controlling soil respiration in a temperate mixed hardwood forest. *Glob. Change Biol.* **4**, 217–227, https://doi.org/10.1046/j.1365-2486.1998.00128.x (1998).

64. Fang, C. & Moncrieff, J. B. The dependence of soil CO₂ efflux on temperature. *Soil Biol. Biochem.* **33**, 155–165, https://doi.org/10.1016/S0038-0717(00)00125-5 (2001).

65. Cable, J. M. *et al.* The temperature responses of soil respiration in deserts: a seven desert synthesis. *Biogeochemistry* **103**, 71–90, https://doi.org/10.1007/s10533-010-9448-z (2011).

Author contributions

J.-B. X. oversaw the study and outlined the M.S.; Z.-Y.W. conducted the experiment and drafted the M.S.; Y.-G. W. & Y.L. involved in selecting soils and polishing the M.S.

Competing interests

The authors declare no competing interests.

Additional information

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