Biotic and Abiotic Contribution to Diurnal Soil CO₂ Fluxes from Saline/Alkaline Soils

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As the second largest carbon flux in terrestrial ecosystems, the soil CO₂ flux is closely related to the atmospheric CO₂ concentration. The soil CO₂ flux is the sum of biotic respiration and abiotic geochemical CO₂ exchange; however, little is known about abiotic CO₂ fluxes in arid areas. To investigate the relative contribution of abiotic and biotic soil CO₂ fluxes over a diurnal course, the abiotic CO₂ 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 CO₂ was a significant component of the soil CO₂ flux in both saline and alkaline soil, which accounted for more than 56% of the diurnal soil CO₂ flux. (2) There was a dramatic difference in the temperature response between biotic and abiotic CO₂ fluxes: the response curves of biotic respiration were exponential in the saline soil and quadratic in the alkaline soil, while the abiotic CO₂ 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 CO₂ flux, biotic respiration was severely underestimated (directly measured soil CO₂ flux was only one-seventh of the biotic CO₂ flux in saline soil, and even an order of magnitude lower in alkaline soil). In addition, the soil CO₂ 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 CO₂ 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 CO₂ flux and its temperature dependence in arid or saline/alkaline land. Further investigations of abiotic CO₂ 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 CO₂ flux between terrestrial ecosystems and atmosphere after photosynthesis¹, the soil CO₂ flux (Fₛ, the exchange rate of CO₂ from soil to atmosphere) is estimated to be 68 ± 4 Pg C.y⁻¹ globally². A small change in Fₛ can significantly alter the atmospheric CO₂ concentration³ and potentially amplify global warming⁴–⁶. Biotic soil CO₂ flux (F₉, 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 forests³. Abiotic soil CO₂ flux (Fₐ, originating from soil abiotic processes such as carbonate weathering and CO₂ dissolution) is reported to be no more than 3–4 gC.m⁻².y⁻¹⁷. Thus, it is customary to assume that Fₐ is purely of biotic origin¹⁸ and equal to the soil respiration rate¹⁰–¹². In this context, most researchers tended to neglect Fₐ in Fₛ studies (annually or on shorter time scale)⁷,¹³,¹⁴ and focused only on soil biotic processes¹⁵,¹⁶.

However, recent studies reported anomalous fluxes or negative soil CO₂ fluxes that cannot be explained by any biotic processes of soil respiration (Rₛ)¹⁷–²¹. In contrast to the marginal contribution assumption, they demonstrated that Fₐ could significantly alter the temporal variation of Fₛ²²,²³. Fₐ might temporally dominate the terrestrial-atmosphere carbon exchange and contribute 19–68% to annual net ecosystem exchange (NEE) in semiarid shrubland²⁴, and even created a sink larger than 100 gC.m⁻².yr⁻¹ in desert regions²⁵–²⁷. Namely, when conditions meet, Fₐ can account for a significant portion of Fₛ: up to 13% in calcareous Mojave Desert soils²⁸, 40% in a Mediterranean region under dry soil conditions²⁹, and more than 75% in the Dry Valleys of Antarctica²⁹. However, only a few studies distinguished Fₛ from Fₐ, and accurate estimates of Fₐ 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 \(F_e\) remains poorly understood, which has resulted in an intensive debate on its magnitude and mechanisms\(^{30,21,23-31}\) and thus induced biases in estimation of soil biotic processes\(^{28,20,34}\). Thus, reliable partitioning of \(F_a\) and \(F_b\) is of critical importance in quantifying their contribution to \(F_e\). As soil biotic and abiotic processes may respond to different drivers and thus respond differently to climate change\(^{34}\), this partitioning is essential in understanding the feedback of the soil carbon cycle in response to climate change\(^{23,28,30}\).

Anomalous fluxes or abiotic CO\(_2\) fluxes were mainly reported in saline or alkaline soil of arid and semiarid land\(^{20,21,23}\), which may occupy 50% of the total land surfaces by the end of this century\(^{30}\). Due the imminent transition to a warmer and more arid climate\(^{39}\), 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 \(F_e\) 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 \(F_e\) 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 \(F_b\) 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 Tienshan 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* Lede. (average canopy cover 17%). Other herbaceous species include *Salsola nitraria* Pall., *Suaeda aequina* 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 Solonetzes) 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 \(F_b\), 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.
Soil analysis. To determine the change in the soil properties following sterilization, soil samples were collected from the soil drums after the CO2 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 K2Cr2O7–H2SO4 Walkley-Black oxidation method. SIC was determined using a modified pressure transducer method.

Data analysis and statistics. Based on the potential sources of CO2, we assumed $F_c$ is the combination of $F_a$ and $F_b$:

$$F_c = F_a + F_b$$  \hspace{1cm} (1)

$F_c$ was determined by measuring the CO2 flux of the control soil (unsterilized soil), representing the total CO2 exchange rate from soil to atmosphere. $F_a$ was determined by measuring the CO2 flux of sterilized soil, representing the CO2 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, 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, or modified van’t Hoff equation - using $F_0$ and $Q_{10}$ as operand which was also equivalent to van’t Hoff equation. 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. 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 CO2 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 CO2 flux and temperature data were used to establish the relationships between soil temperature and $F_c$ 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. Therefore, a mixed...
effects model was used to compare the diurnal patterns of \( F_a, F_b \) and \( F_c \). 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 \( F_a, F_b \) and \( F_c \).

The mean hourly CO\(_2\) fluxes (and their standard errors) of the control and sterilized soils revealed stable diurnal variations of \( F_a \) and \( F_b \) (Fig. 2). \( F_a \) 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 \( F_a \) preceded that of soil temperature by 3 h. In the saline soil, the diurnal amplitude of \( F_a \) was 3.4 \( \mu \text{mol.m}^{-2}.\text{s}^{-1} \) (–0.97 to 2.43 \( \mu \text{mol.m}^{-2}.\text{s}^{-1} \)), whereas in the alkaline soil, the diurnal amplitude of \( F_a \) was 1.58 \( \mu \text{mol.m}^{-2}.\text{s}^{-1} \) (–0.51 to 1.07 \( \mu \text{mol.m}^{-2}.\text{s}^{-1} \)). Over the diurnal course, \( F_a \) 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 \( F_a \) measured in the quartz sand fluctuated around zero throughout the day (Fig. 2C).

Similar to \( F_a \), the unimodal diurnal pattern of \( F_b \) 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 \( F_b \) occurred 2 h earlier compared with soil temperature in both the saline and alkaline soils. The diurnal amplitude of \( F_b \) was 2.18 \( \mu \text{mol.m}^{-2}.\text{s}^{-1} \) (–1.34 to 0.84 \( \mu \text{mol.m}^{-2}.\text{s}^{-1} \)) in the saline soil and 1.42 \( \mu \text{mol.m}^{-2}.\text{s}^{-1} \) (–0.82 to 0.60 \( \mu \text{mol.m}^{-2}.\text{s}^{-1} \)) in the alkaline soil. \( F_b \) 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 \( F_b \) to \( F_a \) 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 \( F_a \) 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 \( \mu \text{mol.m}^{-2}.\text{s}^{-1} \) at 10:00 h and the minimum was 0.35 \( \mu \text{mol.m}^{-2}.\text{s}^{-1} \) 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 \( \mu \text{mol.m}^{-2}.\text{s}^{-1} \)). 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 \( \mu \text{mol.m}^{-2}.\text{s}^{-1} \)), \( F_b \) decreased throughout the night as the soil temperature decreased and reached a minimum value of 0.14 \( \mu \text{mol.m}^{-2}.\text{s}^{-1} \) at 7:00 h. The diurnal amplitude of \( F_b \) was 1.59 \( \mu \text{mol.m}^{-2}.\text{s}^{-1} \) in the saline soil and 0.54 \( \mu \text{mol.m}^{-2}.\text{s}^{-1} \) in the alkaline soil.

The daily averages of \( F_a, F_b \) and \( F_c \) 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 \( F_a \) was 0.10 \( \mu \text{mol.m}^{-2}.\text{s}^{-1} \) in the saline soil, which was only slightly higher than the 0.00 \( \mu \text{mol.m}^{-2}.\text{s}^{-1} \) value for the alkaline soil. The average daily \( F_a \) in the saline soil was –0.63 \( \mu \text{mol.m}^{-2}.\text{s}^{-1} \), which was approximately 170% greater than the value measured in the alkaline soil (–0.37 \( \mu \text{mol.m}^{-2}.\text{s}^{-1} \)). 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 \( \mu \text{mol.m}^{-2}.\text{s}^{-1} \), 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 \( F_a \) on \( F_c \), it 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 \( F_a, F_b \) and \( F_c \).

The correlations between \( F_a \) 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 \( F_a \) in the saline and alkaline soils, respectively. Similarly, soil temperature explained 88% and 65% of the variation in \( F_b \) in the saline and alkaline soils, respectively. Furthermore, according to the

| Soil type and treatment | pH (1:5) | EC (1:5) | SWC (%) | SOC (%) | SIC (%) |
|-------------------------|---------|---------|---------|---------|---------|
| Control saline soil     | 8.86 ± 0.04⁴ | 10.33 ± 0.04⁴ | 3.36 ± 0.05⁵ | 1.32 ± 0.04⁴ | 1.08 ± 0.02² |
| Sterilized saline soil  | 8.85 ± 0.05⁴ | 10.37 ± 0.05⁵ | 3.34 ± 0.06⁷ | 1.31 ± 0.02⁴ | 1.10 ± 0.01¹ |
| Control alkaline soil   | 10.32 ± 0.01⁸ | 1.96 ± 0.02² | 0.94 ± 0.04⁷ | 0.26 ± 0.02² | 0.94 ± 0.01⁴ |
| Sterilized alkaline soil| 10.31 ± 0.02² | 1.95 ± 0.01³ | 0.96 ± 0.05⁹ | 0.27 ± 0.02² | 0.94 ± 0.01⁴ |

Table 1. Soil properties of the control and sterilized saline and alkaline soils. Values represent the mean ± SE (n = 6). Different letters indicate significant differences (\( p < 0.05 \)) between soil type and treatments based on Student’s t-tests.
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 ± 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 ± 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 ± SE, $n = 6$. * and ** denote significant differences between the soils at $\alpha = 0.05$ and $\alpha = 0.01$, respectively.
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 CO2 fluxes reported in the current study (Figs. 2 and 4) are comparable in magnitude to soil CO2 flux measured in situ in the same undisturbed saline desert region27 and in alkaline soil30, suggesting that soil sampling procedures have little effect on soil CO2 fluxes. Previous studies showed that air-drying and sieving of soil could significantly affect soil CO2 fluxes by changing the soil water content, soil structure and soil organic matter fractions35,36. 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 measurements57. We obtained continuous and high frequency measurements of abiotic CO2 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 CO2 in the soil air, and after sterilization, soil biological activity was eliminated and thus stopped CO2 release into the soil. In this case, the change of soil CO2 concentration might lead to the overestimation or underestimation of the abiotic CO2 flux (shifted towards values that are more or less positive). In original soils, positive respiration means its CO2 concentration was the most likely mechanism underlying the abiotic CO2 flux. Although the current study used air-dried soils (Table 1), the abiotic CO2 flux was positively correlated with temperature (bidirectional - CO2 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 CO2 dissolution in soil water22,44,81. The zero CO2 flux measured in the quartz sand and the comparable soil CO2 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 gashphne could not be the main contributor of the abiotic CO2 flux. Moreover, the abiotic CO2 flux was positively correlated with temperature (bidirectional - CO2 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 CO2 dissolution in soil water conform to these characteristics, the CO2 flux resulting from carbonate weathering is relatively small20,34,62 compared with our data. In contrast, the DIC (dissolved inorganic carbon) derived from soil CO2 dissolution is quite large and comparable to daily soil CO2 flux54,55. The modeled CO2 dissolution process produced CO2 flux values that were comparable to our results54. Thus, CO2 dissolution was the most likely mechanism underlying the abiotic CO2 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 CO2 dissolution ability34. 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 CO2 that was absorbed by/emitted from the soil during the diurnal courses (Fig. 3), direct measurements of the soil CO2 flux can be considerably different from the biological respiration rate in saline/alkaline soils. The biotic CO2 flux we obtained was positive throughout the day (Fig. 3), which confirms the nature of biological respiration (unidirectional release of CO2 to the atmosphere)69. Moreover, the higher biotic CO2 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 CO2 flux–temperature relationships (Fig. 6) agreed well with previous reports49,63,64. The 28°C optimum temperature for soil respiration was also comparable to synthesis studies conducted in seven deserts65.

Our results showed that, on daily average, the soil CO2 flux was only one-seventh of the biotic CO2 flux in saline soil, and even an order of magnitude lower in alkaline soil, due to the negative value of the abiotic CO2 flux (Fig. 4). Hence, the soil CO2 flux severely underestimated the biotic CO2 contribution. This result agreed well with a study in a semiarid soil where biological respiration was 3.8 times higher than the measured soil CO2 flux20,34 but was different to a study in Antarctic dry valley soils where the soil CO2 flux measurements overestimated the biotic CO2 flux80. In addition, the temperature response curves of the soil CO2 flux and the biotic CO2 flux were dramatically different. While the soil CO2 fluxes were linearly correlated with soil temperature (Fig. 5), the biotic CO2 flux–temperature relationships were exponential in the saline soil (Fig. 6A) and quadratic in the alkaline soil (Fig. 6B). Accordingly, using soil CO2 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 drought54. 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 CO2 flux would decrease with considerable increase in the soil biology activity. Moreover, the average daily soil CO2 flux in the saline soil was only slightly higher than in the alkaline soil, while the biotic CO2 flux in the saline soil was approximately 2-times higher than in the
alkaline soil (Fig. 4), indicating the soil CO₂ flux measurements also underestimated soil biological activity differences between the saline and alkaline soils.

In general, the soil CO₂ 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 SO₂ flux, so that estimates of the temperature response of soil CO₂ flux and the dynamics of soil organic matter in arid soils are not obscured or underestimated by soil CO₂ flux measurements due to co-varying soil abiotic processes.

Conclusions

The current study is an attempt to distinguish abiotic contribution from the soil CO₂ 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 CO₂ flux in both the saline and alkaline soils. If taken the directly measured soil CO₂ 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 CO₂ fluxes suggested that the responses of the soil CO₂ 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 CO₂ flux and its temperature dependence in arid or saline/alkaline land.

Data availability

All relevant data is contained within the manuscript.

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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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