Soil carbon dioxide emissions from the Mojave desert: Isotopic evidence for a carbonate source

Fiona M. Soper1, Carmody K. McCalley2, Kimberlee Sparks1, and Jed P. Sparks1

1Department of Ecology and Evolutionary Biology, Cornell University, Ithaca, New York, USA, 2School of Life Sciences, Rochester Institute of Technology, Rochester, New York, USA

Abstract Arid soils represent a substantial carbonate pool and may participate in surface-atmosphere CO₂ exchange via a diel cycle of carbonate dissolution and exsolution. We used a Keeling plot approach to determine the substrate δ¹³C of CO₂ emitted from carbonate-dominated soils in the Mojave desert and found evidence for a nonrespiratory source that increased with surface temperature. In dry soils at 25–30°C, the CO₂ substrate had δ¹³C values of −19.4 ± 4.2‰, indicative of respiration of organic material (soil organic matter = −23.1 ± 0.8‰). CO₂ flux increased with temperature; maximum fluxes occurred above 60°C, where δ¹³CO₂ substrate (−7.2‰ ± 2.8‰) approached soil carbonate values (0.2 ± 0.2‰). In wet soils, CO₂ emissions were not temperature dependent, and δ¹³CO₂ substrate was lower in vegetated soils with higher flux rates, higher organic C content, and potential root respiration. These data provide the first direct evidence of CO₂ emissions from alkaline desert soils derived from an abiotic source and that diurnal emission patterns are strongly driven by surface temperature.

1. Introduction

Arid and semiarid landscapes comprise ~40% of the Earth’s terrestrial surface and are expected to be particularly susceptible to global change. As a result, there is considerable interest in the mechanisms controlling surface-atmosphere carbon dioxide (CO₂) exchange in arid lands. Recent research has demonstrated that in arid ecosystems with alkaline soils derived from calcareous parent material, abiotic (nonrespiratory) CO₂ exchange may be an overlooked aspect of the C cycle [Serrano-Oritz et al., 2010]. Although ignoring this flux may lead to overestimates of biological respiration, whether or not this process actually represents a meaningful C sink is currently debated [Schlesinger, 2016]. Several lines of evidence suggest that ecosystems including the saline Gubantonggut desert in western China [Xie et al., 2009], Chihuahuan desert shrubland [Hamerlynck et al., 2013], and unvegetated playa in Nevada [Yates et al., 2012] experience CO₂ uptake and loss apparently mediated by dissolution and exsolution from the soil carbonate pool. However, thus far no in situ measurements have directly identified that carbonates are the source of these emissions. Arid soils are also known to represent a major global carbonate reservoir [Eswaran et al., 2000], and carbonate concentrations in soil can exceed organic C by an order of magnitude or more [Stevenson and Verburg, 2006]. Given that this flux can apparently be moderated by environmental variables such as atmospheric CO₂ concentrations, temperature, and rainfall distribution [Hamerlynck et al., 2013] and that these variables are highly susceptible to change under future climate, detailed description of the process and mechanisms is a pressing priority.

One line of evidence for an abiotic C flux comes from eddy covariance and chamber methods [Xie et al., 2009; Yates et al., 2012; Hamerlynck et al., 2013; Ma et al., 2013], indicating that several arid ecosystems appear to take up CO₂ at night and exhibit positive fluxes from dry soils during the day. These observations are supported by lab experiments showing uptake, and some subsequent release, of isotopically labeled CO₂ [Liu et al., 2015a; Fa et al., 2016a], as well as diel changes in CO₂ concentration [Yates et al., 2012; Liu et al., 2015b; Fa et al., 2016b].

Under dry conditions, the proposed mechanism for these fluxes has some uncertainties [Fa et al., 2016b] but seems to be driven by a diel cycle in which changes in soil temperature shift the CO₂-carbonate-bicarbonate equilibrium, with added effects of turbulence and gas expansion driving convection [Hamerlynck et al., 2013; Fa et al., 2016b]. Dissolution of CO₂ in the soil solution follows Henry’s law, being temperature dependent and proportional to the partial pressure of CO₂ in the soil [Serrano-Oritz et al., 2010]. At night, cooling of dry soils drives upward air migration and deposition of water vapor, along with...
CO₂ solubility is greater at lower nighttime temperatures, although the extent to which dissolution occurs is dependent on the availability of soil water [Fa et al., 2016b]. During the day as surface temperatures increase, carbonate solubility in water films decreases and the expansion of gases plus surface turbulence decreases the concentration of CO₂ in the soil pore spaces, driving outgassing of CO₂ from soil water films and causing CaCO₃ to again precipitate [Roland et al., 2013; Fa et al., 2016b]. Lab experiments confirm that this process is highly dependent on temperature [Chevallier et al., 2016] as well as the rate of change of temperature [Liu et al., 2015b]. Under field conditions, surface CO₂ fluxes show a positive correlation with temperature up to 55°C in the Chihuahuan desert [Hamerlynck et al., 2013].

Carbon stable isotopes (δ¹³C) offer a useful way to resolve CO₂ sources, using Keeling plots to identify the δ¹³CO₂ of the evolved fluxes, and are most effective when the potential sources have well-separated isotopic values [Pataki et al., 2003]. Soil CO₂ fluxes derived from biological respiration (plant roots or microbes) would be expected to closely reflect the highly depleted δ¹³C values generated by photosynthesis (e.g., −22 to −24.5‰ for the arid woody shrub Larrea tridentata [Rundel and Sharifi, 1993]) or present in soil organic carbon, with comparatively little fractionation [ŠantRůčková et al., 2000; Werth and Kuzyakov, 2010]. Solid soil carbonates in the Mojave desert, conversely, have much more enriched δ¹³C values, close to −1‰ [Stevenson and Verburg, 2006], and CO₂ in ambient air is close to −8‰. This significant separation, combined with estimates of fractionation during carbonate dissolution and exsolution, can be used to partition the sources of soil CO₂ fluxes. While lab incubations suggest that CO₂ evolved from these soils can have enriched values close to that of carbonates [Stevenson and Verburg, 2006], no evidence from the Mojave (or any arid lands, to our knowledge) exists to identify carbonates as a source in situ.

The Mojave desert is an arid ecosystem occupying almost 124,000 km², characterized by alkaline, carbonate-dominated soils, and strong diel variations in temperature. Previous observations from Mojave show that CO₂ fluxes from dry soils are highest in summer during the middle of the day, where high soil surface temperatures (up to 80°C) are unlikely to support microbial activity (personal observation). In this study, we used in situ measurements of the C isotope composition of the soil surface CO₂ flux to test the hypothesis that CO₂ emissions from dry soils originate from soil carbonates rather than biological respiration. We measured soil CO₂ fluxes at a range of surface temperatures, under both dry and wet conditions, and used Keeling plots to compare the δ¹³CO₂ with the isotopic composition of potential substrates.

2. Materials and Methods

2.1. Site

Sampling was conducted at the Nevada Desert Free-Air Carbon dioxide Enrichment (FACE) Facility (Nye County, Nevada; 36°49'N, 115°55'W; elevation 970 m), 90 km northwest of Las Vegas, Nevada, in June 2009. The site is contained within the Nevada test site (U.S. Department of Energy) and has been protected from recreation and grazing disturbances at least 55 years [Jordan et al., 1999]. Mean annual rainfall is 140 mm, occurring primarily as winter storms and also as brief, localized summer events. Air temperatures at the site range from −10°C (winter minimum) to 47°C (summer maximum). Diel temperature variation is large, and in summer, midday soil surface temperatures can reach >75°C [McCcalley and Sparks, 2009]. Vegetation at the site is classified as L. tridentata–Ambrosia dumosa–Lycium spp. desert scrub, with vegetation cover of ~20%. Soils are classed as Aridsols, derived from calcareous alluvium, with carbonate values reported in the range of ~0.025 g C cm⁻² at 0–10 cm [Marion et al., 2008]. Surface soils (A1 horizon to 16 cm) are loamy sands (89% sand, 6% silt, and 5% clay), with heterogeneous gravel content to 1 m and a pH of 8–9 [Jordan et al., 1999]. Subsoils are well drained and lack a caliche layer.

2.2. Soil and Gas Sampling

Ten soil cores were taken to a depth of 5 cm beneath L. tridentata and plant interspaces. All soil samples were passed through a 2 mm screen to remove large rock fragments and dried at 60°C for 72 h. Soils were
homogenized using a mortar and pestle. Samples were analyzed for $\delta^{13}$C and %C using a continuous flow isotope ratio mass spectrometer (IR-MS; Model V Advantage; Thermo Scientific, Bremen, Germany). A replicate set of samples was analyzed for carbonate $\delta^{13}$C using a Thermo Finnigan GasBench II (Thermo, San Jose, USA) coupled to the IR-MS using the phosphoric acid-calcium carbonate reaction method outlined in Revez et al. [2001]. All isotope analyses were conducted at the Cornell University Stable Isotope Laboratory.

Field CO$\textsubscript{2}$ sampling was conducted using static chambers [McCalley et al., 2011]. Soil collars (25.5 cm in diameter and 15 cm tall) were installed approximately 7.5 cm into the soil profile in June 2004. Collars were sealed with gastight transparent lids fitted with rubber septa and vents. Temperature at the soil surface was monitored over the course of each incubation using a fine-wire thermocouple inserted 1 cm into the soil, with values recorded every 1 min. Time of day when measurements were taken was staggered to provide a range of soil surface temperatures (from ~08:00 A.M. to 07:00 P.M.). Headspace CO$\textsubscript{2}$ samples were measured under either dry (ambient) or postwetting conditions (18 h after the application of 10 mm water). A precipitation event of 2 mm occurred 18 days prior to the beginning of sampling (Western Regional Climate Centre). Soil water content for this site in June averages 2–3% volumetric water content [Nowak, 2012]. Headspace samples were taken every 30 min for 2 h (postwetting) or 3 h (dry soils). Samples were withdrawn using a gastight syringe fitted with a 21-gauge needle (BD, Franklin Lakes, USA). This needle gauge is large enough to avoid isotopic fractionation of CO$\textsubscript{2}$ during filling (Sparks and McCalley, unpublished data). Samples were injected into evacuated 20 mL glass vials fitted with butyl rubber septa and transported to Cornell University for analysis. Travel standards were employed to check for any vial leakage. The $\delta^{13}$C-CO$\textsubscript{2}$ and CO$\textsubscript{2}$ concentrations were measured on a GasBench II coupled to the IR-MS described above.

### 2.3. Statistical Analysis

Keeling plot intercepts for sample CO$\textsubscript{2}$ concentration and $\delta^{13}$C were solved using Model II (standard major axis) regression, which has been determined to be more suitable for Keeling data than simple linear regression in reducing errors when extrapolating far from the data [Pataki et al., 2003]. This analysis was performed using the “Imodel2” package in R [Legendre, 2014]. Several samples (all from dry sampling) were excluded from analyses if Keeling plot linearity was poor (generally $r^2$ less than 0.35) and slopes were nonsignificant. CO$\textsubscript{2}$ flux rates were determined using linear regression applied to concentration increase over time for the first 2 h of measurement. Linear regression was used to analyze the relationship between temperature and CO$\textsubscript{2}$ flux and source composition. Student’s t test was used to compare CO$\textsubscript{2}$ fluxes between cover types.

### 3. Results

The $\delta^{13}$C values of soil organic C and soil carbonate C at the Mojave site were well separated, averaging $-22.8\%$ to $-23.4\%$ and $0.1\%$ to $0.3\%$, respectively (Table 1).

In dry soils, $\delta^{13}$C of the CO$\textsubscript{2}$ source (Keeling plot solution) moved away from the organic C value and toward the carbonate C value as soil temperature increased up from 33°C to an average of 75°C (Figures 1 and 2). This linear relationship was significant for both interspace and *Larrea* soils ($P < 0.008$ and $P < 0.0287$, respectively; slope of the two lines not significantly different) and the two combined ($\delta^{13}$C = 2.14 × temp + 77.6, $r^2 = 0.63$, $P < 0.0001$; Figure 2). This strong relationship was not seen in postwetting soils; where the $\delta^{13}$C of the CO$\textsubscript{2}$ source varied by less than 5% within each vegetation type (Figures 1 and 2), although there was a significant relationship between source $\delta^{13}$C and temperature for *Larrea* soils ($P < 0.0005$). Temperatures in postwetting soils did not exceed 55°C. In wet soils, the CO$\textsubscript{2}$ source value was on average 4.4‰ higher for interspace than *Larrea* soils, which contained a greater percentage of depleted organic C (Table 1).

| Table 1. The $\delta^{13}$C of Soil (Organic and Carbonate Fractions) Beneath *Larrea tridentata* or Plant Interspace (0–5 cm Depth)*a* | Soil %C (Carbonate) | Soil %C (Organic) | $\delta^{13}$C Organic | $\delta^{13}$C Carbonate |
| --- | --- | --- | --- | --- |
| Interspace | 2.15 ± 0.36 | 0.19 ± 0.04 | $-22.8 \pm 1.1$ | 0.3 ± 0.1 |
| *Larrea tridentata* | 2.48 ± 0.60 | 0.55 ± 0.06 | $-23.4 \pm 0.2$ | 0.1 ± 0.2 |

*a*Values are average ± 1 SD (n = 5 per cover type).
There was a positive linear relationship between soil CO$_2$ flux from dry soils and soil temperature (°C) for both interspace and Larrea soils (P < 0.0007 and P < 0.0017, respectively; slope of the two lines not significantly different) and for the combined cover types (P < 0.0001, flux = 9.74 × temp / 342.9, $r^2 = 0.71$; Figure 3). Postwetting fluxes were significantly greater (495–2595 mg C m$^{-2}$ h$^{-1}$ wet versus 13–518 mg C m$^{-2}$ h$^{-1}$ dry) but did not show a significant relationship with temperature (Figure 3). CO$_2$ fluxes from postwetting Larrea soils were significantly greater than from interspace soils (P < 0.0088).

4. Discussion

In this study, we found isotopic evidence for a shift toward a carbonate-derived CO$_2$ source under high daytime surface soil temperatures in the carbonate-dominated soils of the Mojave desert. We also found that high surface temperatures (>69°C) support the greatest rates of CO$_2$ efflux from dry soils under summer conditions.
Under wet conditions, it appears that microbial and/or autotrophic respiration increases significantly, leading to greater CO$_2$ fluxes and average $\delta^{13}$CO$_2$ source values closer to that of soil organic C.

While natural abundance isotopic methods have been applied to show abiotic sources for CO$_2$ emissions in Antarctic dry valleys [Shanhun et al., 2012], they have not to our knowledge been applied in situ to characterize CO$_2$ sources. This system is characterized by having two soil C pools (organic C and carbonate C) with distinct and nonoverlapping $\delta^{13}$C values, of $\sim 23\%$ and $\sim 0\%$, respectively. In both Larrea and interspace soils under dry conditions, we found that the $\delta^{13}$C-CO$_2$ source (derived from Keeling plots) increased linearly with surface temperature, from $\sim 22\%$ at 33°C to a maximum of $\sim 5\%$ at 69°C. The high end of this range represents some of the most isotopically enriched C source values reported for CO$_2$ sources from untreated soils [Pataki et al., 2003]. CO$_2$ flux rate also increased linearly over this temperature range. Thus, this pattern is consistent with low fluxes of respiration-derived CO$_2$ (from soil organic matter (SOM) or root respiration) from the soil surface at lower temperatures and an increasing flux of isotopically heavy CO$_2$ derived from carbonate precipitation as temperatures increase. Although this mechanism has been reported thus far only for calcium carbonate, it is possible that nocturnal dissolution of sodium carbonate (also present in Mojave soils at this pH [Wood et al., 2005]) and subsequent diurnal decomposition of sodium bicarbonate at high surface temperatures [Whittig and Janitzky, 1963] also contribute to the abiotic flux.

Although microbial respiration is often thought to increase with temperature, this relationship would be expected to reverse as soils become very hot (especially in very dry soils). Cable et al. [2011], for example, found that soil respiration decreased substantially between 20 and 50°C in the Sonoran desert. This, combined with isotopic evidence, suggests that the high CO$_2$ efflux we see at high daytime soil temperatures does not represent biological respiration but rather an increase in the processes thought to drive carbonate precipitation and thus CO$_2$ release. Specifically, higher soil temperatures reduce the solubility of CO$_2$ and drive convection-assisted CO$_2$ outgassing.

In contrast, rates of CO$_2$ efflux in wet soils were not temperature dependent between 24 and 55°C (wet soils do not typically reach the same high temperatures as dry soils) but were greater in Larrea soils than in interspaces. Water addition is routinely observed to increase biological respiration in arid (noncarbonate) soils [Austin et al., 2004], and thus, this pattern fits with an interpretation of biological respiration being the primary source of CO$_2$ from these soils under wet conditions. The $\delta^{13}$CO$_2$ values are offset from that of SOM (on average 5% enriched for Larrea soils and 9% enriched for interspace soils) and are consistent with lab incubations of Mojave soils under comparable temperature and moisture conditions [Stevenson and Verburg, 2006]. Because biological respiration is expected to have a comparatively small fractionation effect (up to 2.1‰ [Werth and Kuzyakov, 2010]), this offset suggests that in wet soils some diurnally evolved CO$_2$ is still derived from the carbonate source. Alternatively, exchange of atmospheric CO$_2$ (at $\sim 7\%$) with CO$_2$ in the soil pore space (in equilibrium with soil carbonate) as soils dry during the incubation [Schlesinger, 1985] might also contribute to the apparent isotopic enrichment of the
headspace CO₂. The smaller offset for Larrea is consistent with Larrea having greater rates of CO₂ flux, most of which is probably biological in origin and consistent with its larger soil organic C pool and possible autotrophic root respiration.

There is currently a knowledge gap in our understanding of carbon cycling in arid, carbonate-dominated systems. The data presented here add an independent line of isotopic verification to the growing body of evidence that abiotic processes are important element of the C cycle in desert soils. We show that diurnally evolved CO₂ arises in part from carbonate sources, providing a source to balance the nighttime CO₂ uptake observed in arid systems [Hamelryck et al., 2013; Liu et al., 2015b; Yates et al., 2012; Xie et al., 2009] and likely keeping the system at (or close to) a state of C equilibrium [Schlesinger, 2016].

Acknowledgments
We thank the Cornell University Stable Isotope Laboratory and John Pollak for the assistance with sample analysis. We also thank Eric Knight, Stan Smith, Lynn Fenstermaker, and everyone at the Nevada Desert FACE, DOE-NTS, and Bechtel Nevada for their assistance with field logistics and for access to the field site. This research was funded by the National Science Foundation (NSF) award DEB-0237674 to Jed Sparks and by the Cornell NSF Interdisciplinary Graduate Education, Research, and Training Program in Biogeochemistry and Environmental Biocomplexity. Author contributions: J. P. S. formulated the original idea and developed the methodology; J. P. S., C. K. C., and K. L. S. conducted the field sampling; C. K. C. and K. L. S. analyzed the samples; J. P. S., C. C., and F. M. S. analyzed the data and performed the statistical analysis; and F. M. S. and J. P. S. interpreted the data and wrote the manuscript. All original data used in this study are included in Tables S1 and S2 in the supporting information.

References
Austin, A., L. Yahdjian, J. Stark, J. Belnap, A. Porporato, U. Norton, D. Ravetta, and S. Schaeffer (2004), Water pulses and biogeochemical cycles in arid and semiarid ecosystems, Oecologia, 141(2), 221–235, doi:10.1007/s00442-004-1519-1.
Cable, J. M. et al., (2011), The temperature responses of soil respiration in deserts: A seven desert synthesis, Biogeochemistry, 103(1–3), 71–90, doi:10.1007/s10533-010-9448-z.
Chevallier, T., L. Cournac, S. Hamdi, T. Gallali, and M. Bernoux (2016), Temperature dependance of CO₂ emission rates and isotopic signature from a calcareous desert, J. Arid Environ., 135, 132–139, doi:10.1016/j.jaridenv.2016.08.002.
Eswaran, H., P. F. Reich, J. M. Kimble, F. H. Beinroth, E. Padmanabhan, and P. Moecharen (2000), Global carbon sinks, in Global Climate Change and Pedogenic Carbonates, edited by R. Lal et al., pp. 15–25, CRC Press, Boca Raton, Fl.
Fa, K., Z. Liu, Y. Zhang, S. Qin, B. Wu, and J. Liu (2016a), Abiotic carbonate dissolution traps carbon in a semiarid desert, Sci. Rep., 6, 23570–23577, doi:10.1038/srep23570.
Fa, K.-Y., Y.-Q. Zhang, B. Wu, S.-Q. Qin, Z. Liu, and W.-W. She (2016b), Patterns and possible mechanisms of soil CO₂ uptake in sandy soil, Sci. Total Environ., 544, 587–594, doi:10.1016/j.scitotenv.2015.11.163.
Hamelryck, E. P. L. Scott, E. P. Sánchez-Cañete, and G. A. Barron-Gafford (2013), Nocturnal soil CO₂ uptake and its relationship to subsurface soil and ecosystem carbon fluxes in a Chihuahuan desert shrubland, J. Geophys. Res. Biogeosci., 118, 1593–1603, doi:10.1002/jgrb.200495.
Jordán, D. N., S. F. Zitzer, G. R. Hendrey, K. F. Lewin, J. Nagy, R. S. Nowak, S. D. Smith, J. S. Coleman, and J. R. Seemann (1999), Biotic, abiotic and performance aspects of the Nevada Desert Free-Air CO₂ Enrichment (FACE) Facility, Glob Change Biol., 5(6), 659–668, doi:10.1046/j.1365-2486.1999.00255.x.
Legendre, P. (2014), Imodel2: Model II regression. R package version 1.4-2.
Liu et al (2015b), 30, 59, doi:10.1016/j.soilbio.2015.11.028.
M. Rundel, P. W., and M. Shari (1996), Hidden, abiotic CO₂ fluxes in Antarctic soils, J. Arid Environ., 31(3), 321–329, doi:10.1016/j.jaridenv.2016.08.002.
Reeve-Knight, M. E., D. K. Smith, J. O. Parnall, and J. A. Berry (2003), The application and Interpretation of Keeling plots in terrestrial carbon cycle research, Global Biogeochem. Cycles, 17(1), 1022, doi:10.1029/2001GB001850.
R. K. Revez, K. M., M. M. Landwehr, and J. Keybl (2001), Measurement of δ 18O values of CaCO₃ using a Thermomix Finnigan GasBench II Delta Plus XL continuous flow isotope ratio mass spectrometer with application to devils hole cave DH-11 calcite. USGS Open-File Report 01-257.
Roland, M., et al. (2013), Atmospheric turbulence triggers pronounced diel pattern in karst carbonate geochemistry, Biogeosciences, 10(7), 5009–5017, doi:10.5194/bg-10-5009-2013.
Rundel, P. W., and M. Shari (1993), Carbon isotope discrimination and resource availability in the desert shrub Larrea tridentata, in Stable Isotopes and Plant Water Relations, edited by J. R. Ehleringer, A. E. Hall, and G. D. Farquhar, Elsevier, San Diego, Calif.
SantnullChová, H., M. I. Bird, and J. Lloyd (2000), Microbial processes and carbon-isotope fractionation in tropical and temperate grassland soils, Funct. Ecol., 14(1), 108–114, doi:10.1046/j.1365-2435.2000.00402.x.
Schlesinger, W. H. (1985), The formation of caliche in soils of the Mojave desert, California, Geochim. Cosmochim. Acta., 49(10), 57–66, doi:10.1016/0016-7037(85)90191-7.
Schlesinger, W. H. (2016), An evaluation of abiotic carbon sinks in deserts, Global Change Biol., doi:10.1111/gcb.13336, in press.
Serrano-Oritz, P., M. Roland, S. Sanchez-Moral, J. A. Janssens, F. Domingo, Y. Godderis, and A. S. Kowalski (2010), Hidden, abiotic CO₂ fluxes and gaseous reservoirs in the terrestrial carbon cycle: Review and perspectives, Agric. For. Meteorol., 153(3), 321–329, doi:10.1016/j.agrformet.2010.01.002.
Shanahan, F. L., P. C. Almond, T. J. Clough, and C. M. S. Smith (2012), Abiotic processes dominate CO₂ fluxes in Antarctic soils, Soil Biol. Biochem., 53, 99–111, doi:10.1016/j.soilbio.2012.04.027.
Stevenson, B. A., and P. S. J. Verburg (2006), Efluxed CO₂–¹³C from sterilized and unsterilized treatments of a calcareous soil, Soil Biol. Biochem., 38(7), 1727–1733, doi:10.1016/j.soilbio.2005.11.028.
Werth, M., and Y. Kuzyakov (2010), ¹³C fractionation at the root-microorganisms–soil interface: A review and outlook for partitioning studies, Soil Biol. Biochem., 42(9), 1372–1384, doi:10.1016/j.soilbio.2010.04.009.
Whittig, L. D., and P. Jantzke (1963), Mechanisms of formation of sodium carbonate in soils, Eur. J. Soil Sci., 14(2), 322–333, doi:10.1111/j.1365-2389.1963.tb00956.x.
Wood, Y. A., R. C. Graham, and S. G. Wells (2005), Surface control of desert pavement pedologic process and landscape function, Cima Volcanic Field, Mojave desert, California, Catena, 59(2), 205–230, doi:10.1016/j.catena.2004.06.001.

Xie, J., Y. Li, C. Zhai, C. Li, and Z. Lan (2009), CO$_2$ absorption by alkaline soils and its implication to the global carbon cycle, Environ. Geol., 56(5), 953–961, doi:10.1007/s00254-008-1197-0.

Yates, E. L., A. M. Detweiler, L. T. Iraci, B. M. Bebout, C. P. McKay, K. Schiro, E. J. Sheffner, C. A. Kelley, J. M. Tadić, and M. Loewenstein (2012), Assessing the role of alkaline soils on the carbon cycle at a playa site, Environ. Earth Sci., 70(3), 1047–1056, doi:10.1007/s12665-012-194-x.