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Greenhouse Gas Production and Transport in Desert Soils of the Southwestern United States

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Abstract

Deserts comprise a large portion of the Earth's land area, yet their role in the fluxes and cycles of greenhouse gases is poorly known and their likely response to climate change largely unexplored. We report a reconnaissance investigation of the concentrations and fluxes of CO_2, CH_4, and N_2O along two elevation (climate) gradients in the southwestern United States. In-soil concentrations of CO_2 increased with elevation (up to 5,000 ppm). Concentrations of CH_4 declined with depth in all soils (to less than 1 ppm), but the rates of decrease with depth increased with elevation. In contrast, concentrations and depth trends of N_2O varied erratically. Soils were net CO_2 sources (0 to >1,500 kg $CO_2 \cdot ha^{-1} \cdot year^{-1}$), and net CH_4 sinks (0.2 to >3 kg $CH_4 \cdot ha^{-1} \cdot year^{-1}$). The small and variable N_2O fluxes were inconsistent with the trends in soil N $\delta^{15}N$ values, which decreased by 5‰ to 6‰ over about 1,000 m of elevation. The high soil N $\delta^{15}N$ values (up to nearly 17‰ at the lowest elevation) indicate that there is a soil N loss mechanism that is highly depleted in ^{15}N, and gaseous losses—either NH_3 or N_2O/N_2—are suspected of driving these values. In summary, there appears to be a strong climate control on both soil CO_2 and CH_4 concentrations and to a lesser degree on calculated fluxes. The soil N trace gas concentrations indicate that deserts can be either small sources or sinks of N_2O and that there may be significant consumption of arid soil N_2O.

Plain Language Summary

We conducted the first comprehensive study of how greenhouse gas production and consumption varies with climate and season in the deserts of California and Nevada. Desert soils are significant consumers of atmospheric methane, an important greenhouse gas. Soils are both small producers and small consumers of nitrous oxide. Given the small nitrous oxide emission rates, there must be an unidentified gaseous loss of nitrogen occurring in desert soils. Finally, the carbon dioxide emission rates, coupled with the amounts of organic carbon in the soils, suggest that deserts—like many ecosystems—will lose organic carbon due to global warming and will in turn act to accentuate greenhouse gases in the atmosphere.

1 Introduction

Deserts lie at the climatic limits of the Earth's biosphere. These biomes, which make up about a third of the Earth's land area, are particularly sensitive to even small absolute changes in their water balance, and climate changes may drive portions of these landscapes toward large ecological responses, or, in extreme cases, lifelessness (Safriel & Adeel, 2005; Schimel, 2010). The role that these biomes play in the present global biogeochemical system, and their responsiveness to anthropogenically driven global change, is at best imperfectly understood. Recent global-scale studies, however, suggest that rapid carbon turnover and interannual variation in productivity in arid and semiarid biomes accounts for significant portions of the year-to-year change in the Earth's terrestrial carbon sink strength (Ahlström et al., 2015; Poulter et al., 2014). To explore this, we examine three greenhouse gases that are highly impacted by soil processes: CO_2, CH_4, and N_2O.

Here we report the results of two year-long studies of the soil concentrations and fluxes of CO_2, CH_4, and N_2O in the southern Great Basin and Mojave Deserts of the southwestern United States. While just one of several arid regions in the United States, this region adjoins (and shares flora with) the bulk of the Great Basin Desert to the north and the Sonoran Desert to the south and south east. We measured in-soil trace gas concentrations along two climosequences in this region. At these study sites, well-defined ecological zones are systematically associated with changes in elevation and allow one to both understand climate impacts on present processes and use these relationships to project the impacts of changes in climate conditions.

2 Background

The Mojave and southernmost Great Basin Deserts are relatively high-elevation deserts that extend from southeastern California into southern Nevada and into small areas of both Utah and Arizona. The “true” desert receives less than 330 mm of rainfall per year, which encompasses the regions below ~1,500 m above sea level (Thomey et al., 2014). However, due to the enormous elevation differences in this basin and range type of landscape, the desert vegetation transitions systematically from open shrubland flora at low elevations to denser, tree-dominated systems at higher elevations. Here we span observations from the warm and dry lower elevations to the higher-elevation semiarid woodlands of the region.

There is at best a poor understanding of soil CO_2 fluxes in desert systems (Galbally et al., 2008) due to a lack of data. A recent review of soil respiration in global deserts reported complex respiration responses to temperature and moisture, with apparent discontinuities in temperature responses between the warm and colder deserts (Cable et al., 2011). However, data for the Mojave were limited to one location, and it is not possible to extrapolate to adjacent elevations and ecological zones. Reported mean rates of soil respiration for this experiment are on the order of 4,500 kg $CO_2 \cdot ha^{-1} \cdot year^{-1}$, with a range from 0 to 45,000 (Cable et al., 2011).

Soil respiration is the counter-process to plant inputs of organic matter, the long-term balance between them being the storage of C as soil organic matter (Amundson, 2001). Thus, respiration can only be fully evaluated by simultaneously considering the soil C pool size. How the opposing processes of inputs and losses will respond to changing environmental conditions is not well constrained. A recent report based on the effects of CO_2 fertilization effects on net primary production (NPP) suggests that Mojave Desert soils may gain C due to increased inputs (Evans et al., 2014). However, this experiment did not include effects of increasing temperature and changing soil water balances, which are first-order controls on both NPP and microbial respiration (Shen et al., 2009).

Less well studied than the soil CO_2 budgets are the fluxes of the important greenhouse gases methane (CH_4) and nitrous oxide (N_2O). Methane consumption in aerobic soils is driven by a group of soil microorganisms known as methanotrophs (Shulka et al., 2013). Some decades ago, it was recognized that desert soils are significant CH_4 sinks (Striegl et al., 1992), and CH_4 consumption rates from deserts have been estimated to average 1.10 kg $CH_4 \cdot ha^{-1} \cdot year^{-1}$ (Dutaur & Verchot, 2007). However, there have been no studies designed to systematically understand the climate controls on the rates of the CH_4 sink, and to date, only a few data from Asian deserts (e.g., Wang et al., 2005) have been added to the original desert data of Striegl et al. (1992). The climatic and geographical controls on CH_4 consumption in deserts are thus poorly constrained.

Finally, the least understood major greenhouse gas in desert soils is N_2O. Nitrous oxide is produced both during oxidative and reductive steps in soils (Butterbach-Bahl et al., 2013), though the net production tends to be highest under anaerobic conditions. It has long been assumed that the only sink of N_2O is reduction by denitrifying microorganisms (Butterbach-Bahl et al., 2013). However, there is growing evidence of net consumption of N_2O by nitrous oxide-reducing, nondentrifying organisms, some of them aerobic, though the extent to which this is accomplished in deserts is not yet documented (Jones et al., 2013, 2014; Orellana et al., 2014; Sanford et al., 2012).

3 Materials and Methods

3.1 Field and Laboratory

The primary study was conducted in the Trail Canyon area of Fish Lake Valley, NV, on the northeastern edge of the White Mountains (Figure 1; Oerter & Amundson, 2016). Fish Lake Valley lies at the southern margin of the Great Basin Desert, where it transitions into the Mojave Desert to the south. A climosequence of four study sites along an elevation gradient was established (see Table 1 and Oerter & Amundson, 2016, for more information). The vegetation at the lowest-elevation site was dominated by *Atriplex* (saltbush) species, grading to *Coleogyne* (blackbrush) at the next higher-elevation site. As elevation increased along the transect, *Artemisia* *tridentata* (sage brush) became more prevalent, and *Pinus monophyla* (pinyon pine) was the dominant tree at the highest site. The sites chosen have soils formed on Holocene-aged alluvium (Reheis & Block, 2007), though there may be some differences in ages between the sites.

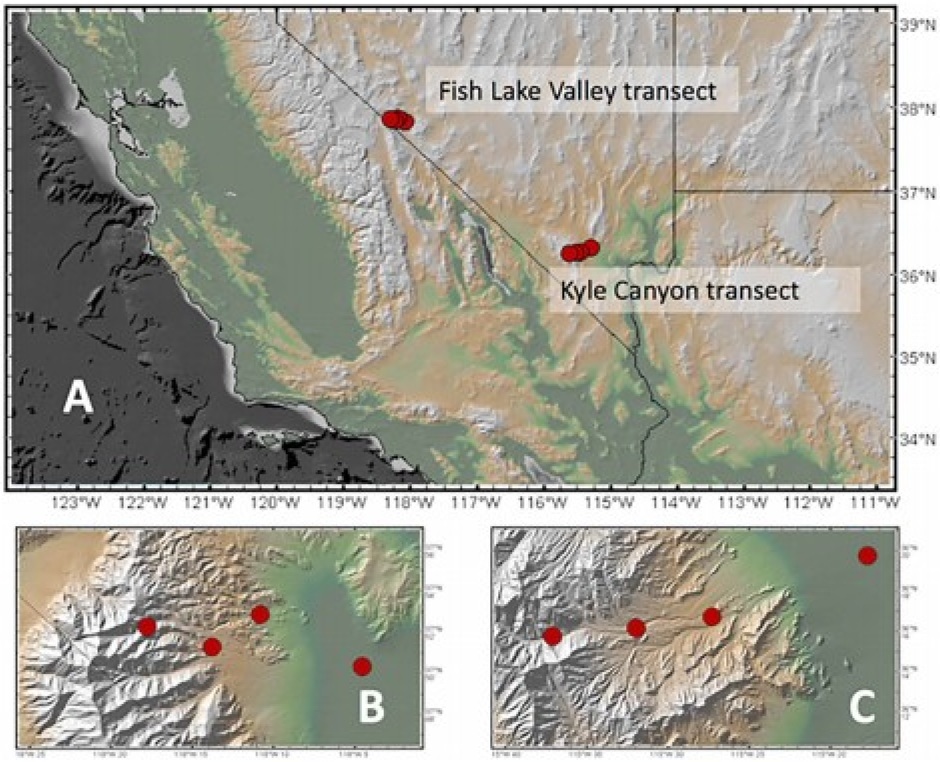

Figure 1. Location of study sites: (a) overview map, (b) Fish Lake Valley, and (c) Kyle Canyon.

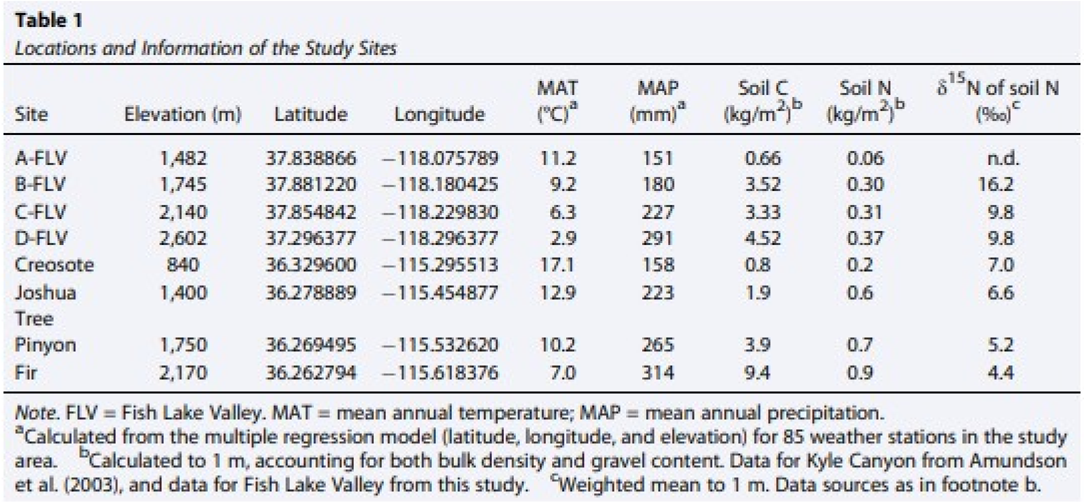

Table 1
Locations and Information of the Study Sites

Site	Elevation (m)	Latitude	Longitude	MAT (°C)[a]	MAP (mm)[a]	Soil C (kg/m^2)[b]	Soil N (kg/m^2)[b]	δ^{15}N of soil N (‰)[c]
A-FLV	1,482	37.838866	−118.075789	11.2	151	0.66	0.06	n.d.
B-FLV	1,745	37.881220	−118.180425	9.2	180	3.52	0.30	16.2
C-FLV	2,140	37.854842	−118.229830	6.3	227	3.33	0.31	9.8
D-FLV	2,602	37.296377	−118.296377	2.9	291	4.52	0.37	9.8
Creosote	840	36.329600	−115.295513	17.1	158	0.8	0.2	7.0
Joshua Tree	1,400	36.278889	−115.454877	12.9	223	1.9	0.6	6.6
Pinyon	1,750	36.269495	−115.532620	10.2	265	3.9	0.7	5.2
Fir	2,170	36.262794	−115.618376	7.0	314	9.4	0.9	4.4

Note. FLV = Fish Lake Valley. MAT = mean annual temperature; MAP = mean annual precipitation.
[a] Calculated from the multiple regression model (latitude, longitude, and elevation) for 85 weather stations in the study area. [b] Calculated to 1 m, accounting for both bulk density and gravel content. Data for Kyle Canyon from Amundson et al. (2003), and data for Fish Lake Valley from this study. [c] Weighted mean to 1 m. Data sources as in footnote b.

At each location in Fish Lake Valley, soils were excavated to at least 1 m, soils were sampled by horizon for subsequent laboratory analyses (Table S1), and instrumentation was installed. Decagon (model EC-TM) volumetric water content and temperature sensors were installed at depths of 10, 25, 50, and 100 cm and connected to data loggers placed in waterproof canisters that were buried at shallow depths (Table S2). Air temperature and relative humidity sensors (Decagon model VP-3) were installed on posts 50 cm above the land surface. Readings from all sensors were recorded on 0.5-hr intervals.

Soil atmosphere samples were obtained from wells constructed from 6.35-mm-outer-diameter, 2-mm-wall, stainless steel tubing with 11-mm-outer-diameter stainless steel drive points welded to the tip. Approximately twenty 2-mm holes were drilled every 5 mm above the tip to create a ~2-cm perforated zone through which gas was collected. At each site, one well was emplaced to each depth of 10, 25, 50, and 100 cm if possible. The two highest-elevation sites, due to high gravel content, required that the deepest wells be driven to the desired depth from a partially excavated pit and the pit was backfilled around the well. The wells were sealed with replaceable rubber septa caps. At the lowest elevation (Site A) two additional sets of wells were installed to assess spatial variability with approximately a 3-m spacing between each set of wells.

In the lab before each field sampling event, 12-ml Exetainer vials were purged with dry N_2 for 10 s at >1-L/min flow rate before being quickly sealed with a dry unpunctured septum screw cap. In the field, the wells were first flushed by drawing 36 ml of air through a Drierite desiccant cartridge into a 60-ml syringe. This was expelled, and another 36 ml was slowly withdrawn and injected into the sample vial while the N_2 was expelled through a 1-cm 22-ga vent needle inserted into the septum. The needles were quickly removed following the gas transfer. Atmospheric air samples were collected a few centimeters above the land surface. Replicate measurements were performed. In total, 180 ml of soil air was withdrawn from each well for each sampling, which represents a sampling radius of 3.5 cm around each well assuming a porosity of 50%. The Fish Lake Valley sites were sampled five separate times: May, August, and November of 2013 and April and May of 2014. During each sampling event, all of the soil gas wells at each site were sampled, and soil water content and temperature, and air temperature data were retrieved from the data loggers.

Within a few days of sampling, the samples of soil and atmospheric air and quality assurance/quality control air (a working gas standard) were analyzed on a Shimadzu GC-14A gas chromatograph with a thermal conductivity detector for determination of CO_2, flame ionization detector for CH_4, and electron capture detector for N_2O samples.

Soil samples from the initial site excavations and instrument installations were air dried and sieved to separate the gravel size fraction from the <2-mm size fraction. Particle size was measured using the hydrometer method (Gee & Bauder, 1986). For soil organic carbon and nitrogen analysis, an aliquot of the <2-mm size fraction was ground to a fine powder using a mortar and pestle. The ground samples were soaked in 0.5-M HCl for 24 hr (1-g soil in 50-ml HCl) and subsequently centrifuged and washed three times with deionized water. Samples were then oven dried at 50 °C. Soil organic C and N content and isotopic composition were measured by continuous-flow dual-isotope analysis using a Elemental Analyzer interfaced to an IsoPrime 100 mass spectrometer at the Center for Stable Isotope Biogeochemistry at the University of California, Berkeley. Carbon and nitrogen isotopic compositions are reported in standard delta notation as per mil (‰) deviations from Pee Dee belemnite and atmospheric N_2, respectively.

To supplement the Fish Lake Valley sites, and to compare them to sites located well within the Mojave Desert, we further examine soil CO_2 concentrations first reported in Amundson, Chadwick, and Sowers (1989). Briefly, eight times over a period of about a year (September 1986 to October 1987), soil CO_2, temperature, and soil moisture were measured at four depths within four key elevation/biotic zones (Tables 1 and S4). Here we use these data, along with additional soil properties data from Amundson, Chadwick, Sowers, and Doner (1989) to calculate the flux of CO_2 for approximately 1 year.

3.2 Calculation of Soil-Atmosphere Fluxes

We used the in-soil trace gas concentrations (c_s), soil physical properties (volumetric water content), and soil textural properties (bulk density and gravel content) at four depths in each soil profile (see Tables S1–S3), along with boundary layer trace gas concentration ($c_s(0)$) at each sampling location, to calculate soil trace gas fluxes. Surface soil-atmosphere gas exchange for each location was estimated using a three-step process:

1. Estimate soil trace gas diffusivity (D_s) at each measurement depth.
2. Calculate a continuous soil profile of trace gas concentration ($c_s(z)$) and diffusivity ($D_s(z)$) for soil depths $0 \leq z \leq 100$ cm using one of two methods (linear segments and polynomial).
3. Calculate the surface trace gas flux to the atmosphere (f_s) from the continuous profiles using integrated profile production/consumption or surface flux methods.

These three steps and their associated methods are described in detail below. All calculations were done using MATLAB (MathWorks, Inc.).

3.3 Soil Gas Diffusivity Coefficients

The flux rates calculated from in-soil gas concentration profiles depends on the soil gas diffusion coefficient. The diffusion coefficient of gases in the atmosphere was set at 0.1381 (Massman, 1998). In soil, diffusivity is lowered due to pore space (solids and soil water reduce air porosity) and increased tortuosity of gas flow. High bulk density, gravel content, and water content all reduce pore space and affect the tortuosity. Numerous methods have been developed to estimate soil gas diffusivity. We selected a model, equation 8 from Moldrup et al. (1999), for which we have data to constrain the calculations:

$$\frac{D_s}{D_o} = \frac{\varepsilon^{2.5}}{\sqrt{\phi}}, \quad (1)$$

where D_o is diffusivity of gas in air, D_s is diffusivity in soil, ε is air-filled porosity (cm^3/cm^3), and ϕ is total soil porosity (cm^3/cm^3). Values for ϕ and ε were calculated as

$$\phi = 1 - \left(\frac{\rho_b}{\rho_p}\right) = \varepsilon + \theta, \quad (2)$$

where ρ_b is bulk density, ρ_p is particle density (assumed at 2.65 g/cm^3), and θ is volumetric water content.

3.4 Soil Gas Concentration and Diffusivity Profile Models

To estimate in-soil gas fluxes and production/consumption rates, for each measurement location, profiles of measured c_i (concentration of gas i) and D_s were used to calculate continuous profiles of these variables. Continuous c_i and D_s profiles ($c_i(z)$ or $D_s(z)$) were fit to the measured profiles using either a linear approach or a second-order polynomial model. The linear approach consisted of fitting line segments between measured c_i values, including the boundary layer value. The second-order polynomial model was fit using least squares and forced through the boundary layer value at $z = 0$. In the case of D_s profiles, the $z = 0$ value was set to be identical to the shallowest measured depth (10 cm).

We encountered variable boundary layer concentrations, especially for CO_2 and N_2O ($c_i(0)$; Figure 3). Thus, for surface flux calculations, we assigned constant boundary condition values of 397, 2.1, and 0.350 ppm for CO_2, CH_4, and N_2O, respectively, which were within the range of commonly measured values. Additionally, for CH_4 and N_2O, we calculated the net production or consumption of the trace gas per soil layer to examine how these processes behave versus soil depth. Summing these values yields an integrated net production/consumption for the soil profile that does not depend on surface boundary condition and thus is a separate assessment of soil processes. These calculations are discussed in the next section.

3.5 Calculation of Soil-Atmosphere Fluxes

Gas exchange between soils and the atmosphere is accomplished, to first order, by molecular diffusion. In this work, we assume steady state conditions. While it is obvious that the fluxes change temporally, the adjustment times of soil gas profiles to perturbations are on the order of a day or so, such that except for large and very recent perturbations, the profiles at least approach steady state conditions. Thus, the flux (f_i) of gas across a soil layer can be approximated using Fick's first law as the differential equation:

$$f_i = D_\mathrm{s}\frac{\partial c_i}{\partial z}. \quad (3)$$

The time dependence of a gas *i* with concentration c_i at depth below the soil surface $0 \le z \le 100$ cm is a combination of concentration change via diffusion (Fick's second law) and the local production term, expressed as

$$\frac{\partial c_i}{\partial t} = D_\mathrm{s}\frac{\partial^2 c_i}{\partial z^2} + p_i, \quad (4)$$

where D_s is the diffusion coefficient for the gas corrected for water saturation in the soil and p_i is the local production/consumption term. A value of $p_i > 0$ indicates net production of trace gas *i* by the soil. Assuming the trace gas profile is at steady state, equation 4 can be set to zero, and p_i can be calculated directly from the gas concentration profile with depth using an inverse model:

$$p_i = -D_\mathrm{s}\frac{\partial^2 c_i}{\partial z^2} = -D_\mathrm{s}\frac{c_i(z+\mathrm{d}z) - 2c_i(z) + c_i(z-\mathrm{d}z)}{\partial z^2}. \quad (5)$$

The depth-integrated production/consumption term $p_i(z)$ gives the net soil trace gas flux out of the soil ($f_{i\mathrm{net}}$),

$$f_{i\mathrm{net}} = \int_0^z p_i \mathrm{d}z. \quad (6)$$

Second, the soil gas surface flux was calculated at $z = 0$ using Fick's first law. For a steady state profile, the depth-integrated $f_{i\mathrm{net}}$ should equal the soil gas surface-atmosphere flux:

$$f_i^{z=0} = D_s^{z=0}\frac{c_i(\mathrm{d}z) - c_i(0)}{\partial z}. \quad (7)$$

3.6 Model Validation and Limitations

As reviewed by Wolf et al. (2011), numerous studies have used soil profile concentrations to calculate the soil-to-atmosphere fluxes of CO_2, CH_4, and N_2O. This method is not without limitations or important caveats. These limitations include nonsteady state situations due to the soil response to a perturbation, uncertainties in the calculation of the diffusion coefficient, and the fact that the method misses rapid near-surface processes that may drive gas fluxes in particularly dry environments. However, it also has been shown that this method provides flux rates comparable to chamber measurements for CO_2 (Tang et al., 2003; Wolf et al., 2011). As for N_2O, there has been less agreement on how well the calculated fluxes match chamber measurements. Wolf et al. (2011) showed that calculated fluxes are lower than chamber measurements, while Pihlatie et al. (2007) found the reverse. To further examine this, we used a large published data set of soil profile N_2O concentrations and chamber flux measurements (Van Greongen et al., 2005). Using the published soil water contents, and the diffusion model described above, we found a good agreement between the two approaches (Figure S1). In this paper, with its emphasis on the reconnaissance nature of soil trace gas research in deserts, we consider the flux calculations to be largely comparable to chamber measurements, but note that the relatively low number of measurements restricts estimates of annual or seasonal fluxes.

4 Results and Discussion

4.1 Soil Climate

The soil climatic conditions varied systematically both temporally and spatially (Figure 2). Soil temperatures decreased with depth in the summer, increased with depth in winter, and decreased with elevation. Temperatures reached nearly 40 °C at 10-cm depth at the lowest elevation in the summer and were below freezing at >50 cm for several months during the winter at the highest elevation. Soil volumetric moisture content varied between values of 0.015 and 0.14 cm^3/cm^3 (Tables S2 and S3). Overall soil moisture content increased with elevation; the duration and infiltration depth of precipitation events also increased with elevation.

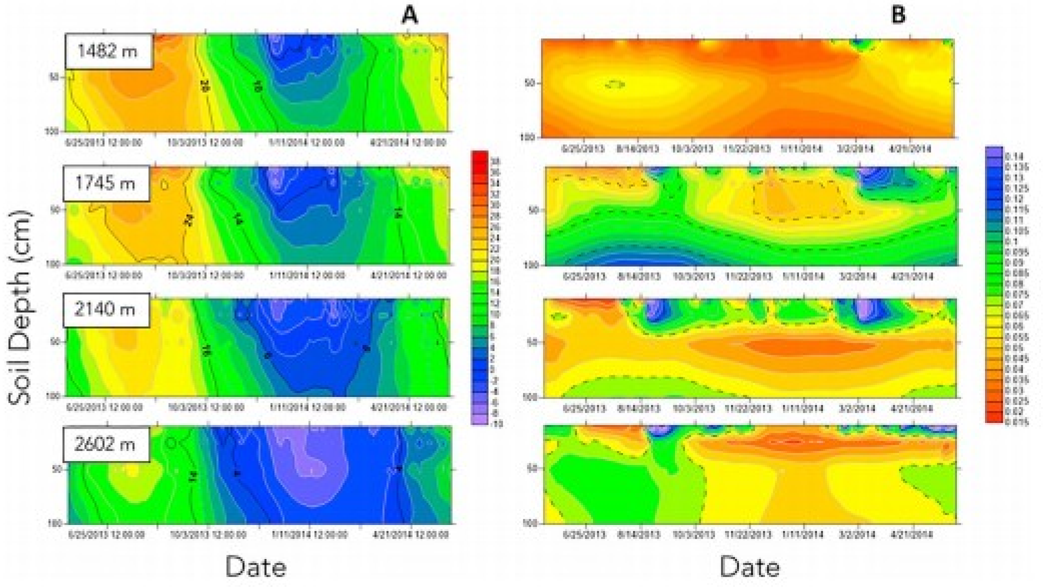

Figure 2. Variation in (a) soil temperature (°C) and (b) soil volumetric water content (cm^3/cm^3) versus soil depth and time. The panels are arranged from low (top) to high (bottom) elevation.

4.2 Soil CO_2

Soil CO_2 concentrations generally increased with soil depth, reflecting the effects of production coupled to diffusional transport toward the overlying atmosphere (Figures 3a–3d). Concentrations at any given depth increased with elevation, reaching concentrations up to 5,000 ppm (Figure 3d). Winters were periods of low CO_2 at all elevations (<1,000 ppm at all sites except the highest elevation), while CO_2 increased with depth and elevation during the summer months, reflecting a combined effect of temperature and available soil moisture on soil respiration. The previously published soil

CO_2 concentrations at Kyle Canyon, NV, also showed the same trends (Amundson, Chadwick, & Sowers, 1989).

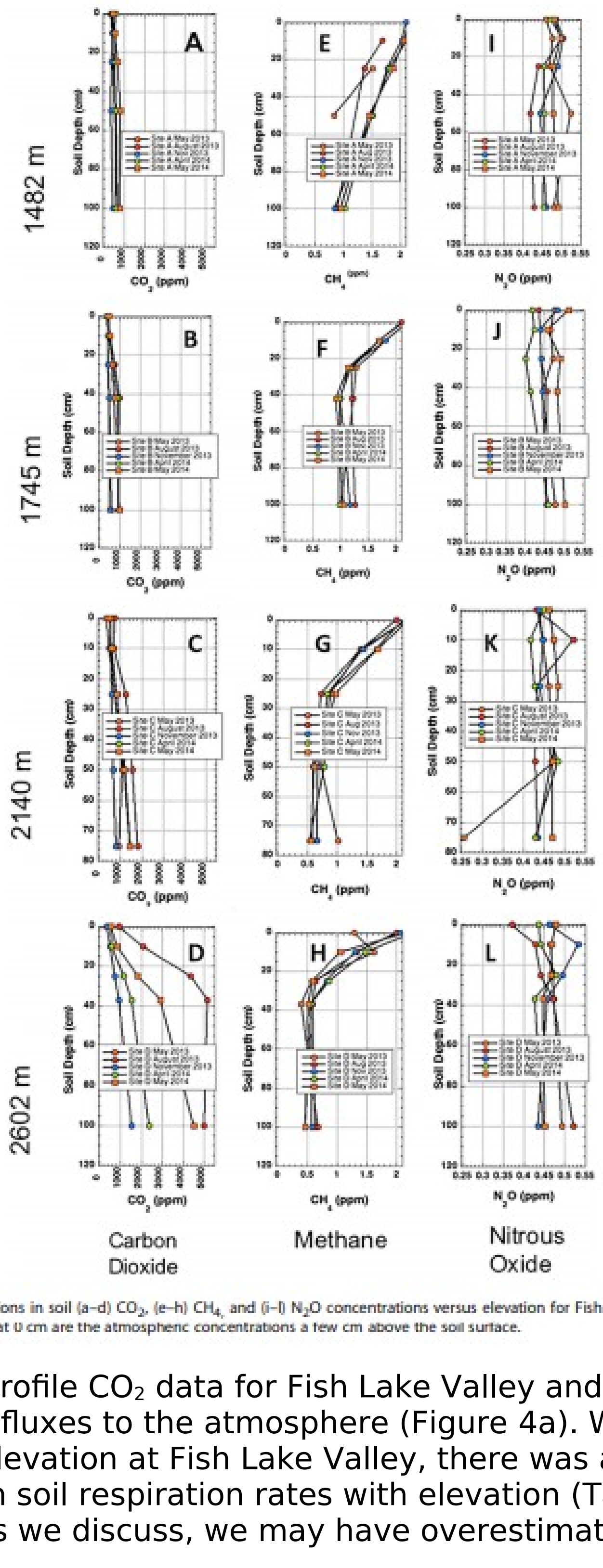

Figure 3. Variations in soil (a–d) CO_2, (e–h) CH_4, and (i–l) N_2O concentrations versus elevation for Fish Lake Valley, NV. Concentrations at 0 cm are the atmospheric concentrations a few cm above the soil surface.

The soil profile CO_2 data for Fish Lake Valley and Kyle Canyon was used to calculate fluxes to the atmosphere (Figure 4a). With the exception of the highest elevation at Fish Lake Valley, there was a general trend of increasing maximum soil respiration rates with elevation (Table 2). As with the other gas fluxes we discuss, we may have overestimated the effect of gravel on soil diffusivity; thus, our flux estimates at the highest elevation may be minimum values.

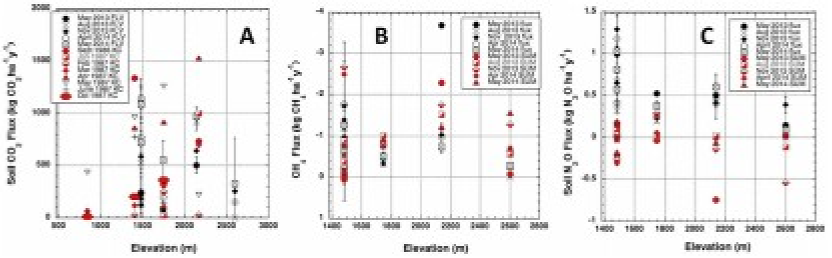

Figure 4. The calculated fluxes of (a) CO_2, (b) CH_4, and (c) N_2O for Fish Lake Valley, NV (and for CO_2, Kyle Canyon, NV). For CO_2 fluxes, rates across the soil-atmosphere boundary only are reported. For CH_4 and N_2O, both the calculated flux across the soil-atmosphere boundary ("flux") and the sum of soil net production/consumption ("sum") are reported.

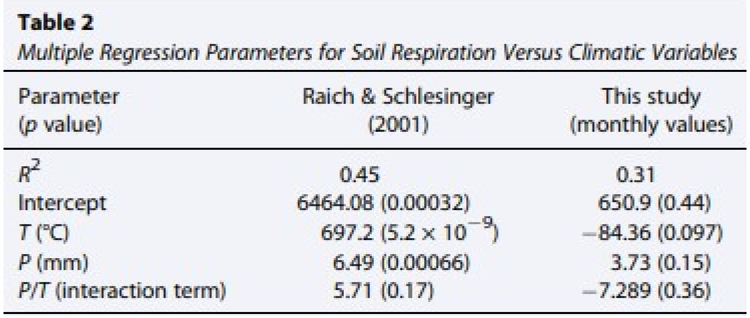

Table 2

Multiple Regression Parameters for Soil Respiration Versus Climatic Variables

Parameter (*p* value)	Raich & Schlesinger (2001)	This study (monthly values)
R^2	0.45	0.31
Intercept	6464.08 (0.00032)	650.9 (0.44)
T (°C)	697.2 (5.2×10^{-9})	−84.36 (0.097)
P (mm)	6.49 (0.00066)	3.73 (0.15)
P/T (interaction term)	5.71 (0.17)	−7.289 (0.36)

There are few other data on soil respiration for the Mojave or Great Basin Desert. A review by Cable et al. (2011) reports only one location for the Mojave: a free air CO_2 enrichment study (Schaeffer et al., 2003). The respiration rates for the CO_2 enrichment study (about 4,500 kg $CO_2 \cdot ha^{-1} \cdot year^{-1}$; Cable et al., 2011) appear to be higher than the respiration measurements made here. An alternative approach is to compare soil respiration to NPP. We assume that soil respiration rates are approximately 0.5 times that of annual NPP (soil respiration is roughly 50% living root respiration and about 50% soil organic matter decomposition that must equal NPP at steady state), though the actual value may be less in low-productivity ecosystems such as the Mojave (Bond-Lamberty et al., 2004; Hanson et al., 2000). In a long-term study of NPP in Rock Valley, NV (just north of Kyle Canyon), Lane et al. (1984) found that average NPP was 301 kg dry matter per hectare, which converted to C (and CO_2) is 550 kg $CO_2 \cdot ha^{-1} \cdot year^{-1}$ (the range by year varied from 330 to 1052 kg $CO_2 \cdot ha^{-1} \cdot year^{-1}$). Thus, the fluxes from CO_2 profiles appear to closely match the observed NPP values at the low elevations.

It is essential to recognize that soil respiration rate is strongly dependent on the soil C content (substrate), as well as climate. Thus, in order to better constrain the effects of soil temperature and moisture on CO_2 emission, we normalized the calculated respiration rates (Figure 4a) by the soil C content (Table 1), giving kg CO_2 per unit time per kilogram of soil C (essentially an estimate of the decomposition constant). A simple analysis of how this rate varies with soil temperature and soil water content is shown in Figure 5. For both sites, normalized respiration tends to increase with temperature and decrease with moisture (p < 0.05 for Kyle Canyon and <0.10 for Fish Lake Valley for the temperature regression in Figure 5a). At both sites, as soil temperature increased, soil moisture declined (Figure 5c). It appears that as long as some moisture is available, temperature exerts a strong control on rates of biological processes across all sites. While the linear relationship in Figure 5a is noisy, it implies a Q_{10} for respiration per kilogram of C of 1.6 to 1.7 for both Fish Lake Valley and Kyle Canyon, a temperature response of biological soil C decomposition consistent with those observed in global analyses of decomposition (Koven et al., 2017). Thus, these preliminary measurements suggest that desert soil C, like that in other regions, is likely sensitive to the anthropogenic warming that is ongoing.

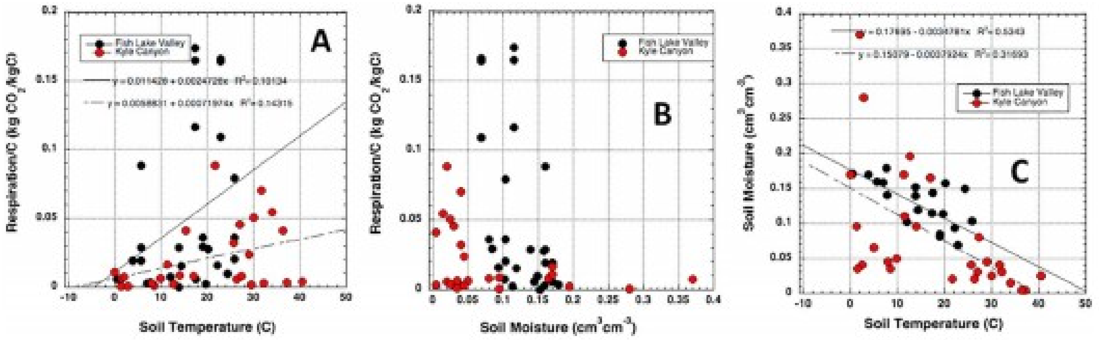

Figure 5. The calculated soil respiration rates for Fish Lake Valley and Kyle Canyon (normalized to the total soil C) versus (a) measured soil temperature and versus (b) soil moisture. (c) The relation between soil temperature and soil moisture.

4.3 Soil Methane

Methane is consumed in aerobic soil due to its oxidation by a group of organisms called methanotrophs. The methanotrophs responsible for the oxidation of atmospheric levels of CH_4 are called “high affinity” groups and have not been cultured in the lab (Shulka et al., 2013). The impact of climatic conditions on the rates of methane oxidation is still uncertain. There is likely a parabolic response of methane oxidation to soil water content (low rates at both dry and wet extremes), while the temperature response has been found to be rather unpredictable based on a review of recent literature (Shulka et al., 2013). Striegl et al. (1992) were the first to discover that Mojave soils are atmospheric methane sinks and that the strength of the sink appears to increase with increasing soil moisture.

Methane concentrations in the soils at Fish Lake Valley all generally decline below atmospheric values with increasing depth, and the rate of decline with depth increases with increasing elevation (Figures 3e–3h). Qualitatively, this suggests that consumption also parallels concentration trends. Figure 4b illustrates the calculated rates of consumption versus elevation and time of year for both soil/atmosphere exchange and the sum of the soil profile net production/consumption rates, which are illustrated in Figure 6. For both surface flux and summation calculations, the apparent trends with elevation are similar and show an increase in CH_4 consumption with elevation up to the third site and then a decline at the highest elevation. The strongest apparent sink rates are in the upper two layers of the soil at all the sites, which are also the zones of highest carbon and, presumably, highest microbial activity. As with CO_2, we suspect that the calculated diffusivities in the very gravelly high-elevation soil may be too low, and thus, fluxes (which have a strong concentration gradient versus depth, as seen in Figure 3h) may underestimate the rates of consumption.

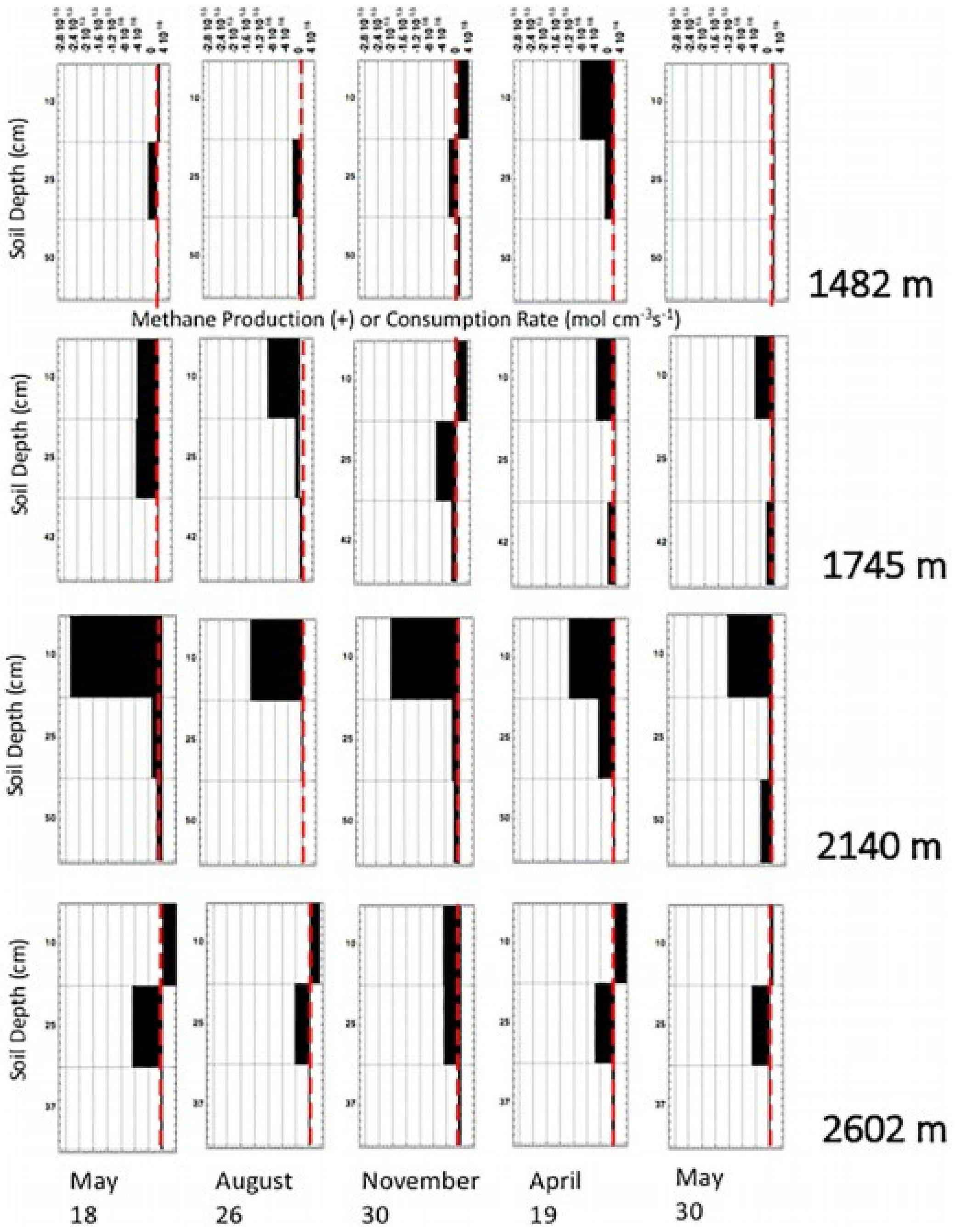

Figure 6. Depth trends in CH_4 consumption (negative) or production (positive) rates versus depth for the soils at Fish Lake Valley, NV.

The methane consumption rates at Fish Lake Valley vary from essentially 0 to about 3.5 kg CH_4·ha·year^{-1} (Figure 7). These rates are similar to those observed by Blankinship et al. (2010) near Flagstaff, AZ. In AZ, where precipitation rates are higher, consumption rates varied from ~3 to ~6 kg CH_4·ha·year^{-1}, about a factor of 2 to 3 greater than those measured at Fish Lake Valley. For Fish Lake Valley, the mean of all samplings is

1.0 kg CH_4·ha·year^{-1}; this consumption rate spread over the total areal extent of deserts (11.24×10^6 km^2) implies a global consumption about 1.1 Tg of CH_4 per year, similar to the 1.5 Tg estimated by Zhuang et al. (2013). While the Fish Lake Valley data are too sparse to provide any improvements to estimates of the global desert soil CH_4 consumption, they do point to a possible climate-driven control on the rates, which should be tested by further observations and research.

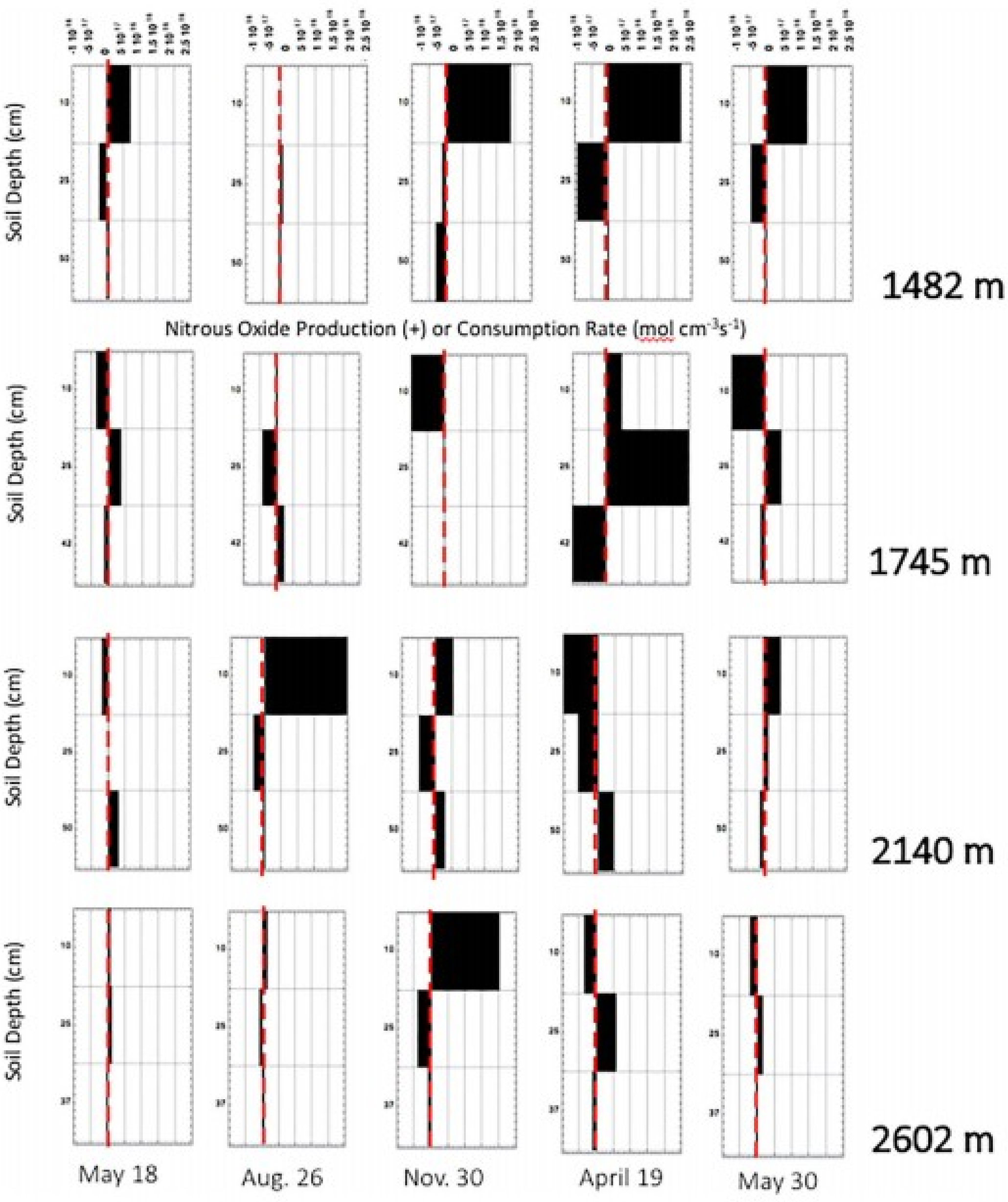

Figure 7. Depth trends in N_2O consumption (negative) or production (positive) versus depth for the soils at Fish Lake Valley, NV.

4.4 Soil Nitrous Oxide

Nitrous oxide production in soils occurs during aerobic (NH_4^+ oxidation to NO_3^-) and anaerobic segments (NO_3^- reduction) of the soil N cycle. While isolated anaerobic sites may seasonally exist even in arid soils, the Fish Lake Valley soils must be largely aerobic due to the low rainfall and soil moisture contents.

One of the surprising observations is that the concentration of N_2O in the soils showed little systematic variation with depth, time, or elevation (Figures 3i–3l). Both the CO_2 and CH_4 concentration profiles qualitatively show variations with season and elevation (Figures 3a–3h), while the N_2O concentrations were largely insensitive. In this section, we explore two hypotheses: (1) that there is little N_2O production or (2) that N_2O is being produced in response to climate, but gross consumption is masking the rates of gross production.

Again, as for CH_4, N_2O fluxes were calculated across the soil-air interface and as the summation of net production/consumption by depth, which is also illustrated in Figure 7. Soil-to-atmosphere fluxes of N_2O were slightly higher (roughly 0.5 kg $N_2O \cdot ha^{-1} \cdot year^{-1}$) than profile summation. The summation calculations produced more apparent net N_2O consumption events. In general, N_2O exchange rates varied from −0.5 to 1.3 kg $N_2O \cdot ha^{-1} \cdot year^{-1}$, with no apparent sensitivity to elevation. The mean N_2O flux for all sites for the surface flux method was 0.5 kg $N_2O \cdot ha^{-1} \cdot year^{-1}$, and for the summation method it was −0.04 kg $N_2O \cdot ha^{-1} \cdot year^{-1}$. The positive flux rates bracket previous NO_2 flux studies in southwestern deserts (Billings et al., 2002; Guilbault & Matthias, 1998), and the negative rates provide additional evidence of dry season consumption of N_2O (Schlesinger, 2013). For example, Matson et al. (1991) observed N_2O consumption rates in an arid steppe of Wyoming of up to −0.5 kg $N_2O \cdot ha^{-1} \cdot year^{-1}$. In the Sonoran Desert of Arizona, Guilbault and Matthias (1998) observed sinks of up to −1.0 kg $N_2O \cdot ha^{-1} \cdot year^{-1}$. Conen and Neftel (2007) speculate that half of the world's soils are N_2O sinks. While the controls on N_2O consumption are variable and not well understood (Chapuis-Lardy et al., 2007), our results provide supporting evidence that gross N_2O consumption in soil may occur under aerobic conditions.

Recent research in molecular biology has demonstrated that the soil N_2O cycle is highly modular and that numerous previously unrecognized organisms are capable of consuming N_2O (Jones et al., 2013, 2014; Orellana et al., 2014; Sanford et al., 2012). These organisms possess the *nosZ* gene that encodes N_2O reductase but lack the genes for nitrate or nitrite reductase (and thus the production of N_2O). This suggests that even in largely aerobic soils, organisms capable of N_2O reduction may exist and that measured N_2O concentrations and fluxes may be the net result of simultaneous production and consumption processes.

The form and relative importance of soil N losses is reflected isotopically in the remaining soil N pool (Amundson et al., 2003; Houlton & Bai, 2009;

Houlton et al., 2015). Gaseous forms of N (N_2O, N_2, or NH_3) all are depleted in ^{15}N relative to the remaining soil organic N pool. Soil N pools, in the absences of major disturbance, are near steady state, with inputs = outputs. The present rates of N deposition in the arid west are low. Data from the National Atmospheric Deposition Program (http://nadp.sws.uiuc.edu) for their collection station outside of Las Vegas show that from 1985 to 2002, the average NH_4, NO_3, and total N was 0.53 (±0.29), 2.51 (±0.98), and 0.98 (±0.42) $kg \cdot ha^{-1} \cdot year^{-1}$, values that are similar to those extrapolated to remote areas across California and Nevada. While these rates must include N deposition from automotive emissions, they are still low. At both Kyle Canyon and Fish Lake Valley, biological N fixing plants are rare or absent, and N fixation must be small relative to other inputs. We assume relatively constant N inputs with elevation, which implies N residence times decline with decreasing elevation and increasing temperature, though we acknowledge deposition rates might increase with precipitation rates.

However, the annual N loss mechanisms for these soils are unclear. Soil leaching is low due to the dry conditions. The total available precipitation (monthly rainfall that exceeds evaporation for that month) increases from 20 to about 200 mm/year at Kyle Canyon (Amundson, Chadwick, & Sowers, 1989). At the lowest elevation, the low amount of available water (20 mm) is reflected by the nitrates and chlorides found in this soil below 15 cm (Figure 9a; Amundson, Chadwick, Sowers, & Doner, 1989). Walvord et al. (2003), through deeper drilling into the vadose zone, found that in the Mojave Desert (at Yucca Mountain, north of Kyle Canyon), nitrates and chlorides accumulated below the rooting zone (estimated by them to be 1 m) in quantities that suggested an annual downward flux rate of 0.09 to 1.17 kg N/year over the past 10,000 years, which is less than or similar to present atmospheric N deposition rates. As illustrated in Figure 9b, the NO_3^--rich zones in this soil have $\delta^{15}N$ values of 5‰ to 7‰, which are enriched relative to atmospheric deposition values that are 0‰ ± 5‰ globally (Amundson et al., 2003) but must be similar along the small transect at Kyle Canyon. Thus, while a considerable fraction of the total soil N output appears to be in the form of NO_3^-, the remaining N loss mechanisms must be considerably more depleted in ^{15}N to drive total soil $\delta^{15}N$ values to these positive values relative to the inputs. There is a decrease in total soil $\delta^{15}N$ values with elevation (Figure 8a), indicating that there is a corresponding decrease in the fraction of the total soil N that is lost by relatively ^{15}N-depleted sources with decreasing aridity.

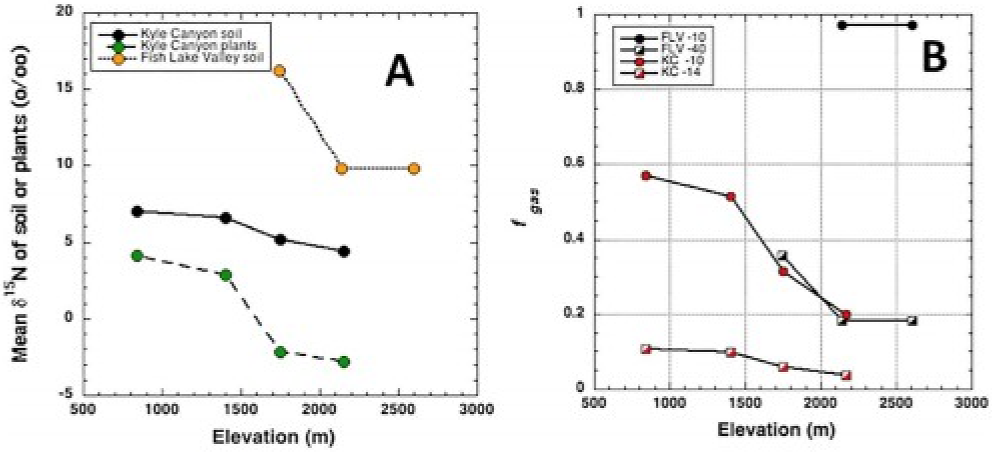

Figure 8. Soil and plant d^{15}N values versus (a) elevation for Kyle Canyon (Amundson et al., 2003) and Fish Lake Valley and (b) the estimated fraction of gaseous N loss (*f*) for the same sites, using equation (8) in the text.

Houlton and Bai (2009) introduced a simple mass balance model that can be used to estimate the fraction (f_{gas}) of total soil N loss that is gaseous. In their global analyses, they assumed gaseous soil N loss was N_2, resulting from denitrification, but all soil N gaseous losses tend to cluster into the same broad isotope effect. The model for calculating the fraction of gaseous loss from soil is

$$f_{\text{gas}} = \frac{\delta^{15}N_{\text{soil}} - \delta^{15}N_{\text{atm}} + \varepsilon_{\text{nitrate}}}{\varepsilon_{\text{nitrate}} - \varepsilon_{\text{gas}}} \qquad (8)$$

where $\varepsilon_{nitrate}$ and ε_{gas} are the isotope fractionation effects of nitrate leaching and gaseous losses, respectively (relative to the soil N pool). Houlton and Bai (2009) used a $\varepsilon_{nitrate} = -0.8‰$. Here we use a more conservative value of −3.0‰ (assigning more possible isotope fractionation effects to nitrate loss; see Figure 9b). Following Mnich and Houlton (2016), we use ε_{gas} values of −10‰ and −40‰ to explore ranges in possible gaseous loss rates under different isotope effects. We note that the gaseous losses can be N_2O, N_2, and/or possibly NH_4 volatilization. The results, in Figure 8b, suggest that the fraction of total N losses as gas (where $f_{gas} + f_{nitrate\ leaching} = 1$) declines with elevation, with gaseous loss *f* values of 0.6 to 0.04 at Kyle Canyon and higher values at Fish Lake Valley (nearly 1 at the driest site, where we have soil N isotope data, to 0.1 at the wettest). Assuming a very low atmospheric N input rate of about 1 kg N·ha^{-1}·year^{-1}, total gaseous loss of inputs by N_2O emissions would be 4.4 kg N_2O·ha^{-1}·year^{-1}, far higher than any observations. As a consequence, we hypothesize that observable N_2O fluxes largely underrepresent total N gaseous losses from these soils, and that the discrepancy appears to increase with increasing aridity, as Houlton et al. (2015) have calculated on a global scale. Clearly, there is a need for more research on the soil N cycle in desert regions. In particular, the enigmatic isotope evidence for N gaseous losses, in dry locations with minimal or even negative N_2O fluxes, will require additional work to test this initial finding and determine the species of apparent gas loss that drives the soil N isotope trends.

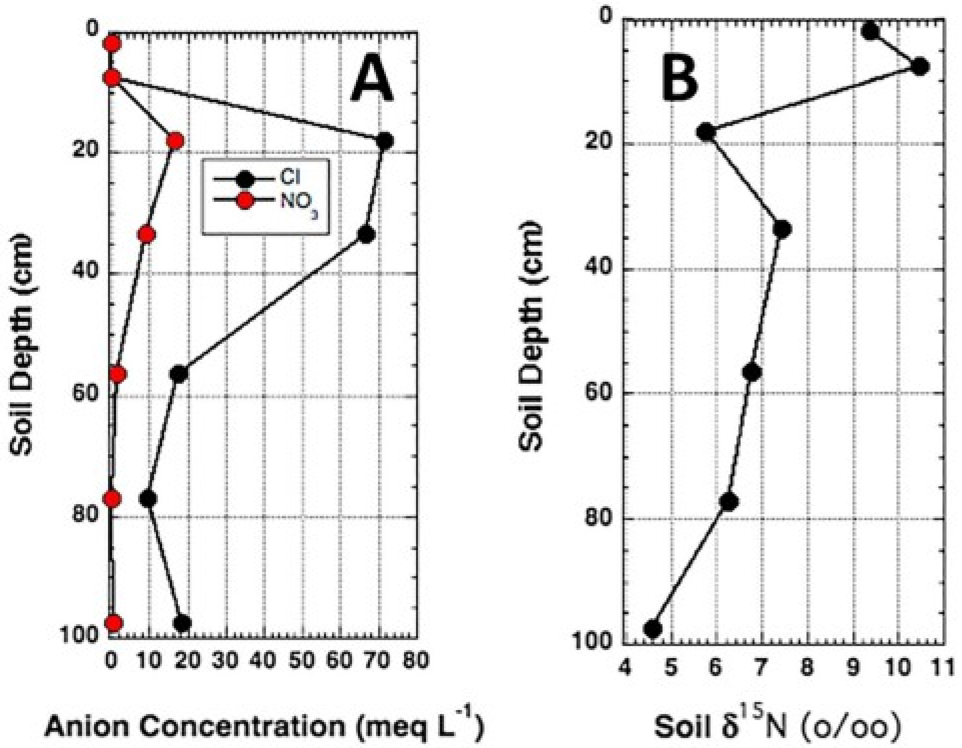

Figure 9. (a) The NO_3 and Cl concentration in saturated extracts for the lowest-elevation site at Kyle Canyon, NV (data from Amundson, Chadwick, & Sowers, 1989), and (b) soil $\delta^{15}N$ values versus depth (Brenner, 1999).

5 Conclusions

This reconnaissance investigation of desert soil trace gases suggests that CO_2 and CH_4 concentrations show systematic changes with depth and elevation, while N_2O does not. We calculated trace gas soil-atmosphere fluxes from in-soil concentration profiles and diffusivities based on soil particle size and water content data, a method that should largely reflect flux measurements made with chambers. Soils are a source of CO_2, a sink of CH_4, and either a small source or sink for N_2O. Methane consumption rates suggest a climate response, which requires additional observations elsewhere. The apparent lack of significant N_2O fluxes from the soils is inconsistent with the soil organic N stable isotope composition, which requires that there are greater losses of ^{15}N-depleted N with increasing aridity, and such losses are likely to be of a gaseous form. In these desert soils, two probable forms of gaseous loss are NH_3 volatilization and/or N_2 produced from the reduction of N_2O.

While these observations provide some guidance into the possible range of values and patterns of trace gas processes in these soils, they also suggest a wealth of research questions that should be addressed before we fully understand arid region trace gas biogeochemistry. Key steps to better understanding these processes, we suggest, include the following: (a) gradient (elevation and climate) studies, (b) using soil depth gas concentrations and isotope measurements, and (c) modeling to determine net exchange and depth-related processes. We also recognize the need for parallel surface flux measurements (using chambers), which were not included in this initial study.

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