1 Coupled soil oxygen and greenhouse gas dynamics under variable hydrology

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3 Karla M. Jarecke¹, Terrance D. Loecke¹, Amy J. Burgin¹

4
5 ¹University of Nebraska-Lincoln, School of Natural Resources, Lincoln, NE 68583-0974

6
7 Corresponding Author:

8 Terrance D. Loecke

9 School of Natural Resources

10 The University of Nebraska—Lincoln, NE

11 616 Hardin Hall

12 3310 Holdrege Street

13 Lincoln, NE 68583-0996

14 E-mail: tloecke2@unl.edu

15 Phone: 402-472-3499

16 Fax: 402-472-2946

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26 **Abstract:** Hot spots and hot moments of greenhouse gas (GHG) fluxes can contribute
27 significantly to overall GHG budgets. Hot spots and hot moments occur when dynamic soil
28 hydrology triggers important shifts in soil biogeochemical and physical processes that control
29 GHG emissions. Soil oxygen (O₂), a direct control on biogenic GHG production (i.e., nitrous
30 oxide-N₂O, carbon dioxide-CO₂ and methane-CH₄), may serve as both an important proxy for
31 determining sudden shifts in subsurface biogenic GHG production, as well as the physical
32 transport of soil GHG to the atmosphere. Recent technological advancements offer opportunities
33 to link *in-situ*, near-continuous measurements of soil O₂ concentration to soil biogeochemical
34 processes and soil gas transport. Using high frequency data, this study asked: Do soil O₂
35 dynamics following short-term (< 8 days) soil saturation correspond to changes in soil GHG
36 concentrations and GHG surface fluxes? We addressed this question in a restored riparian

wetland in Ohio, USA. Changes in subsurface (10 and 20 cm) CO₂ and N₂O concentrations were inversely related to short-term (< 48 h) changes in soil O₂ concentrations. Subsurface CH₄ concentrations, however, did not change in response to soil O₂ dynamics. Changing subsurface GHG concentrations did not necessarily translate into altered surface (soil to atmosphere) GHG fluxes; soil O₂ dynamics at 10 cm did not correspond with changes in surface N₂O and CH₄ fluxes. However, changes in soil O₂ concentration at 10 cm had a significant positive linear relationship with change in surface CO₂ flux. Our study suggests that monitoring near-continuous soil O₂ concentration under dynamic soil hydrology may lead to greater understanding of hot spots and hot moments of GHG emissions. This understanding is increasingly important for estimating the contribution of soil processes to atmospheric GHG concentrations.

Keywords: soil oxygen dynamics, soil moisture, greenhouse gases, biogeochemical processes, soil gas transport

1. Introduction

Brief periods of disproportionately high rates of soil-atmospheric biogenic greenhouse gas exchange (GHGs; carbon dioxide, CO₂, nitrous oxide, N₂O and methane, CH₄), known as “hot moments,” contribute significantly to whole-system GHG budgets (McClain et al., 2003; Groffman et al., 2009; Kuzyakov and Blagodatskaya, 2015). GHG hot moments are often triggered during transitional periods in soil hydrology (e.g., soil re-wetting and thawing) that induce sudden changes in the biological and physical soil environment (Smith et al., 2003; Kim et al., 2012; Malodovskaya et al., 2012). Although hot moments may significantly alter estimates of nutrient fluxes, the biogeochemical and physical mechanisms that drive GHG spatiotemporal variability are poorly understood (Groffman et al., 2009; Vargas et al., 2011). Advancing our mechanistic understanding of soil GHG fluxes requires more accurate understanding of the biological drivers of CO₂, CH₄, and N₂O production/consumption as well as physical GHG transport during soil re-wetting and drying (Chen et al., 2008; Riley et al., 2011; Blagodatsky and Smith, 2012).

An important but understudied biological driver of soil GHG fluxes is soil O₂ availability. Soil O₂ controls redox potential and is the master variable determining aerobic and anaerobic biogeochemical processes, including GHG production and consumption (Burgin et al., 2011; Liptzen et al., 2011; McNicol and Silver, 2014). In lieu of rare direct measurements of soil O₂, scientists commonly extrapolate soil O₂ concentration from the more easily measured soil water content (Heinen et al., 2006; Davidson et al., 2012). Soil moisture predicts soil O₂ concentration under stable conditions, but fails to describe the sudden increase in soil O₂ concentration at thresholds in volumetric soil moisture content (Liptzen et al., 2011; Burgin and Groffman, 2012). Studies that measured soil O₂ as an average concentration in the bulk soil (0-21%) find

concentrations to be dynamic on short temporal scales (hours to weeks) under varying hydrology as a result of seasonal water table or precipitation patterns (Silver et al., 1999; Teh et al., 2005; Liptzen et al., 2011; Burgin and Groffman, 2012; Silver et al., 2012; Hall et al., 2013). In saturated mineral soils, physical resupply of atmospheric O₂ to the soil is limited and the rate of soil O₂ depletion depends on soil temperature (Papendick and Runkles, 1965), suggesting a strong relationship between the rate of soil O₂ depletion and microbial respiration. Dry soils, in contrast, limit biological activity under low solute diffusion and water stress (Harris, 1981), but permit the rapid diffusion of atmospheric O₂ to the soil. When water-filled soil pores begin to drain after prolonged saturation, a sudden shift in the controls on soil gas diffusion (from diffusion through water to air-filled pore spaces) allows the quick re-supply of atmospheric O₂ to the soil. Herein, for simplicity we refer to the quick re-supply of atmospheric O₂ during soil drainage as the “big gulp”. While we are beginning to understand the controls on soil O₂ dynamics, including big gulps, we do not yet understand how the timing of soil O₂ fluctuations correspond to soil-atmospheric GHG fluxes. This shift in diffusivity also applies to gases at higher concentration in the soil than in the atmosphere, e.g., N₂O. For example, rapid influx of atmospheric O₂ to the soil may co-occur with an efflux of soil gases to the atmosphere.

Given the important role of soil O₂ availability on biogeochemical processing, we ask: 1) Does measured soil O₂ correlate to variation in soil GHG concentrations and fluxes? and 2) Can measured soil O₂ act as a physical indicator of soil GHG transport? We predict that short-term soil O₂ fluctuations under varying hydrology may be a useful proxy for understanding the timing of briefly enhanced rates of surface GHG fluxes (hot moments). Using high frequency soil O₂ and soil moisture sensors, we evaluated how soil GHG concentrations and surface GHG fluxes correspond to near-surface (10 and 20 cm depth) soil O₂ dynamics. During sustained soil

saturation, we expected the depletion of soil O₂ and restrictive diffusion of soil GHGs to the atmosphere to result in an increase in subsurface GHG concentrations at 10 and 20 cm and a decrease in GHG surface fluxes. Upon drainage, we expected the rapid influx of atmospheric O₂ (“big gulp”) into soil pore space to co-occur with a decrease in subsurface GHG concentrations and an increase in surface GHG fluxes.

2. Materials and Methods

2.1. Study site

Our research is conducted at the Great Miami Wetland Mitigation Bank, a 46 ha restored wetland in Trotwood, Montgomery County, Ohio, United States (36°46’51” N, 84°20’26” W). Five Rivers Metro Parks, Dayton, OH funded and managed the restoration project, which serves as a wetland mitigation bank to provide compensatory mitigation for impacted waters of the United States, including wetlands and streams, which result from activities authorized under the Clean Water Act (National Research Council, 2001). The restoration involved extensive earth moving during 2011 and planting native wetland and upland vegetation during 2012. Herbaceous plants (a mix of perennial and annual grasses and forbs) dominate the vegetation. The site, previously drained for row crop production for over 100 years, is underlain by poorly drained silty clay loam (Brookston, Fine-loamy, mixed, active, mesic Aeric Endoaqualfs). Average annual regional precipitation is 1005 mm and average annual temperature is 10.8°C. Average daily temperature during our study was 11.93 °C in October 2013 and 21.37°C from June-August 2014. The site received a total 147 mm of rain in October 2013 and approximately 178 mm from June to August 2014.

We installed soil sensors and gas sampling chambers at two locations within the larger site, approximately 200 m apart. We chose the location of the sites to represent differences in soil

texture across the restored wetland (Figure 1 and Table 1). Soil texture at 0-10 cm across the restored wetland was estimated from 120 soil cores collected post-construction in November 2011, summarized as a histogram in Figure 1. The soil at the quick draining (QD) site is sandy clay loam soil at 0-10 and 10-20 cm, which had shorter saturation duration following precipitation events in 2014 compared to the slow draining (SD) site with clay soil (0-10 cm) and clay loam soil (10-20 cm) (Table 1). In September 2013, we installed the QD site with five replicated sampling pits positioned approximately two meters from the nearest neighboring pit. In June 2014, we added the SD site with three replicated sampling pits positioned approximately two meters apart. QD is located in an area where intensive earth moving occurred (2010-2011); SD experienced minimal impact under construction activity.

2.2. Sensor and gas chamber installation

Data loggers at each site (Campbell Scientific CR1000, Nexens 3100-iSIC, and Stevens DOT) recorded soil O₂, soil moisture and soil temperature at 30 minute intervals. We sampled GHG subsurface (10 and 20 cm) concentrations and GHG surface (soil to atmosphere) fluxes during October 2013 at QD and from June to August 2014 at QD and SD. Each pit consisted of two SO-110 soil O₂ sensors (Apogee Instruments, Logan, Utah), two SDI-12 hydra probes (Stevens Water, Portland, Oregon), and four subsurface gas sampling chambers installed horizontally 15-20 cm from a neighboring sensor at 10 and 20 cm depth (Figure 2). The soil O₂ sensor has internal temperature composition circuitry and the data are expressed as relative concentration, i.e., the mixing ratio as a percentage. We used the sensor mV output in open air at 100 % humidity to calibrate individual O₂ sensors relative to the atmosphere. Soil O₂ sensors were secured vertically to a 30.5 cm diffusion head made of perforated polyvinyl chloride (PVC) pipe (1.9 cm ID, 2.54 cm OD). Subsurface gas chambers were constructed with 30.5 cm of

silicone tubing (1.27 cm ID, 0.32 cm wall thickness) secured in perforated PVC pipe (1.9 cm ID, 2.54 cm OD) and fit to a tygon lead that ran to the surface for sampling with a stopcock and syringe. The average time to 95% equilibration with surrounding CO₂ concentration is 9.1 hours for silicone tubing with wall thickness equal to 0.32 cm (DeSutter et al., 2006). Prior to installing soil O₂ diffusion heads and subsurface gas chambers, we drilled a hole of slightly smaller diameter horizontally into the soil wall with a 1.9 cm auger bit. Above each sampling pit, we placed static surface flux chambers (12.5 cm in length x 25 cm inside diameter) outfitted with a PVC lid with a small sampling port needle vent to permit equilibration of internal and external atmospheric pressure (modified from Robertson et al., 1999). We positioned the center of the permanent collars 30 cm from the soil pit wall to ensure surface gas fluxes were collected directly above subsurface gas sampling chambers and soil O₂ diffusion heads. Surface gas chamber collars were inserted 3-5 cm into the soil. We saw no visible evidence of cracking at the soil surface and allowed chambers to settle one week prior to sampling.

2.3. Soil analyses and gas sampling

In June 2014 we collected three replicate 30 cm soil cores at QD and SD. Cores were collected one meter from the surface flux chamber where soils were undisturbed during sensor installation. We cut cores in two sections: 0-10 cm and 10-20 cm. We assessed each section for soil properties including bulk density, percent sand, percent silt, percent clay (hydrometer method, Gee and Bauder, 1986), total carbon, total nitrogen (Dumas Method, Bremner, 1996), and substrate-induced respiration (SIR) as an indicator of microbial activity (West and Sparling, 1986). The two sites were significantly different ($\alpha = 0.05$) in bulk density, texture characteristics, SIR, and total nitrogen, but do not differ in total carbon (Table 1). Average SIR at 0-10 cm was 5 times greater at SD compared to QD soil at 0-10 cm; average SIR at SD near-

surface soil (0-10 cm) was also five times greater than deeper soil (10-20 cm) whereas SIR did not differ between depths at QD (Table 1).

We analyzed soil saturation events at QD in October 2013 (10 events among 5 soil pits) and at QD and SD from June to August 2014 (102 events among 8 soil pits). Soil saturation duration (described in data analysis methods) ranged from 2.5 hours to 8 days. We collected subsurface and surface gas samples 1-2 times prior to precipitation and twice daily immediately following precipitation until soil O₂ returned to stable, near-atmospheric concentration at all locations. In August 2014, we used supplemental irrigation to trigger saturation events by pumping water from an artesian well (approximately 400 meters away from sites) to an elevated, rotating sprinkler that distributed water equally at each pit.

Subsurface gas samples were collected into a 6 mL Exetainer (Labco, Exeter UK) vial previously flushed with N₂ and consisted of 3 mL of chamber soil gas after flushing 1.5 mL from the lead tubing (twice the volume of the lead tubing). Immediately following subsurface soil gas collection, we collected repeated gas samples from surface chamber headspace to measure surface gas flux. Surface gas samples were collected with a stopcock and syringe at regular intervals (0, 15, 30, and 45 minutes) and transferred into a 6 mL Exetainer vial previously flushed with surface chamber air (modified from Parkin & Venterea, 2010). All gas samples were immediately shipped to the University of Nebraska-Lincoln and analyzed on an Agilent gas chromatograph using flame ionized detector for CH₄, electron capture detector for N₂O, and external CO₂ analyzer (LICOR 820). For QA/QC, we analyzed one check sample of known CO₂, N₂O, and CH₄ concentration for every twenty field samples; if the coefficient of variation for checks was less than 3% we accepted GC results that fell within the detection limits of the analytic methods.

2.4. Data analysis: characterizing soil O₂ and saturation events

We characterized rapid increases in soil O₂, also called “big gulps” by our research group, as a sustained increase in soil O₂ concentration that exceeds 1.2% and last for 1 h within a 5 h period. This threshold in soil O₂ concentration was insensitive to diurnal change and sporadic noise in soil O₂ measurements. We quantified the local minimum and maximum soil O₂ concentration within 20 hours before and after the big gulp to understand the magnitude of soil O₂ depletion. To understand how saturation duration influenced soil O₂ depletion, we quantified soil saturation (100 % water-filled pore space) using local threshold values for soil moisture (cm³ cm⁻³). We defined local saturation thresholds by subtracting 0.04 from the 99th percentile of measured soil moisture to remove measurement noise. The start of the saturation event occurred when soil moisture measured above the threshold 4 times in 2 hours; the end of the saturation event occurred when soil moisture measured below the saturation threshold 4 times in 2 hours. These criteria ignored short-term (< 2 hours) increases in soil moisture that did not induce change in soil O₂ concentration. We used the time at the start and end of the saturation event to calculate the total time soil remained saturated (saturation duration) and average soil temperature at each location. We analyzed the response of soil O₂ depletion to saturation duration and average soil temperature during saturation at QD and SD 10 and 20 cm using linear regression models ($\alpha < 0.05$).

We collected GHG samples during saturation and within 24 hours of the big gulp at QD 10 cm (n=16), QD 20 cm (n=8), SD 10 cm (n=13), and SD 20 cm (n=10). To understand the effect of dynamic soil O₂ on measured GHG concentrations and surface fluxes, we estimated: 1) change in subsequent GHG concentration and surface GHG flux for samples collected within 48 hours, and 2) simultaneous change in soil O₂ concentration. We analyzed the linear and quadratic

response of change in subsurface GHG concentration (10 and 20 cm) and surface GHG flux to simultaneous change in soil O₂ concentration when change in soil O₂ was greater than 3%; 3% represents more than two times the average diurnal change in soil O₂ concentration. All data analysis was performed in R© (R Core Team, 2014).

3. Results

During June-August 2014, soil moisture ranged from 0.1 to 0.6 cm⁻³ cm⁻³; soil O₂ concentration ranged from 1.3 to 21% among all QD and SD sampling locations. Following precipitation events in October 2013 and June-August 2014, we observed two repeated patterns in soil O₂ dynamics: 1) Soil O₂ concentration began to decline as soil moisture increased to saturation (Figure 3b solid arrow), and 2) Soil O₂ concentration increased rapidly (defined as “the big gulp”) during soil drainage (Figure 3b dashed arrow). On average, SD (slow draining) had greater decrease in soil O₂ concentration (%) and longer periods of saturation compared to QD (quick draining) from June-August 2014 at both 10 and 20 cm (Table 1). Soil O₂ depletion varied vertically (at 10 and 20 cm) and horizontally (among sampling pits) at SD and QD following June-August 2014 precipitation (Figure 4a-d). Soil O₂ depletion during soil saturation occurred at all sampling locations with the exception of QD “r5” at 20 cm depth (Figure 4). Soil O₂ depletion at all site and depth combinations was a positive linear function of mean temperature during saturation multiplied by saturation duration (Figure 5; $p < 0.001$; $R^2 = 0.41$). Soil O₂ depletion increased significantly ($p < 0.05$) with mean soil temperature $\times$ saturation duration at QD 10 cm ($R^2 = 0.31$, 95% CI; Figure 5a), SD 10 cm ($R^2 = 0.31$, 95% CI; Figure 5b), QD 20 cm ($R^2 = 0.24$, 95% CI; Figure 5c), and SD 20 cm ($R^2 = 0.74$, 95% CI; Figure 5d). Soil

O₂ depletion at SD 20 cm had the strongest linear relationship with soil temperature × saturation duration (Figure 5d).

Soil O₂ dynamics, GHG fluxes, and subsurface GHG concentrations varied considerably between sites SD and QD. Figure 6 contrasts the response of soil O₂ concentration following June precipitation for a single pit at QD and SD. At QD soil O₂ remained relatively stable at 10 and 20 cm following a precipitation event. Simultaneously, CO₂ concentration increases, but N₂O and CH₄ concentration remain relatively unchanged (Figure 6a,c). In contrast, soil O₂ depletion was greater at SD under sustained saturation (1 - 2 days) at 10 and (8 days) 20 cm (Figure 6b,d) and increases in soil CO₂, N₂O, and CH₄ concentrations were greatest at SD 20 cm during soil O₂ depletion to 1.3% (Figure 6d). Surface (soil to atmosphere) CO₂ flux dynamics at QD (Figure 7a) and SD (Figure 7b) were variable. Although CO₂ surface flux was dynamic at both QD and SD, N₂O surface flux increased more at SD than QD whereas CH₄ decreased slightly from pre-saturation conditions at both sites (Figure 7a,b).

Change in GHG concentrations and fluxes are calculated from sequential measurements taken within 48 hours. Change in CO₂ surface flux was a positive linear function of change in soil O₂ concentration at 10 cm (Figure 8a). Change in N₂O and CH₄ surface fluxes were not significantly related to change in soil O₂ concentration at 10 cm (Figure 8b,c). In the subsurface, change in CO₂ concentration at 10 and 20 cm was a negative linear function of change in soil O₂ concentration (Figure 8d $p < 0.001$; $R^2 = 0.58$ at 10 cm and $p < 0.001$, $R^2 = 0.35$ at 20 cm). Change in N₂O concentration at 10 and 20 cm, like CO₂, was a negative linear function of change in soil O₂ concentration (Figure 8e $p = 0.05$; $R^2 = 0.11$ at 10 cm and $p = 0.003$; $R^2 = 0.24$ at 20 cm). CH₄ concentration did not significantly respond to soil O₂ dynamics at 10 cm or 20 cm (Figure 8f).

4. Discussion

Our challenge was to understand if soil O₂ availability corresponds to shifts in subsurface GHG concentrations and surface GHG fluxes. We found that the magnitude of soil O₂ depletion following short-term (hours to days) saturation varied horizontally and with soil depth (10 and 20 cm) and is predicted by duration of soil saturation and average soil temperature during saturation (Figure 5). We detected significant increase in subsurface CO₂ and N₂O concentration in response to saturated soil conditions and soil O₂ depletion (Figure 8d,e). We also determined that rapid soil O₂ increases corresponded with decreases in subsurface CO₂ and N₂O concentration and an increase in surface CO₂ flux (Figure 8a,d,e). We will examine the effect of soil O₂ dynamics on biogenic GHG in greater detail below and explore the implications of soil O₂ dynamics for understanding biogeochemical hot spots and hot moments.

4.1. Soil O₂ effects on biogenic greenhouse gases

CO₂ concentrations had the strongest relationship with soil O₂ dynamics. Soil CO₂ concentration increased when soil O₂ concentrations decreased (Figure 8d), which supports that soil O₂ availability and increased labile substrates under short-term soil saturation creates conditions for soil O₂ consumption and carbon mineralization. Several studies report enhanced CO₂ flux during soil re-wetting (Harris, 1981; Ryals and Silver, 2013) with the magnitude of the CO₂ pulse dependent on the history of drying-rewetting events (Fierer and Schimel, 2002). However, the subsequent decrease in soil CO₂ concentration that we observed co-occurs with big gulps (Figure 8d), a physical process initiated by soil drainage. In addition, big gulps correspond to increased CO₂ surface fluxes while soil O₂ depletion corresponds with decreased CO₂ surface fluxes. This link between short-term, dynamic soil O₂ and CO₂ concentrations as well as near-surface soil O₂ concentration and CO₂ surface flux has not been reported by other studies.

Furthermore, the greatest change in CO₂ concentrations and CO₂ surface flux occurred at SD where magnitude of soil O₂ depletion and subsequent big gulps were greater (Figure 8a,d). Thus, we conclude that greater magnitude of short-term (<48 h) soil O₂ depletion triggers greater shifts in CO₂ concentration and surface CO₂ flux.

Although we found the strongest relationship with soil CO₂ concentration, decomposition rates are mediated by O₂-requiring enzyme activity and are not a direct response to soil O₂ availability (Hall et al., 2014). Enzyme activity may be uninhibited by short-term anaerobiosis and contribute to localized microsites of O₂ consumption (Hall et al., 2014). If rates of CO₂ production remains constant in fluctuating redox soils as suggested by incubation experiments (DeAngelis et al. 2010), subsurface CO₂ concentration and surface CO₂ flux will depend largely on physical processes (e.g., diffusion in water vs. air-filled pore space). Therefore, sampling CO₂ concentration and surface flux at greater frequency may further separate changes in CO₂ production (a biological process) from changes in subsurface CO₂ concentration and surface flux (a physical process).

Unlike CO₂, N₂O is controlled by dynamic production and consumption processes — nitrification and denitrification (Burgin et al., 2010; Burgin and Groffman, 2012; Morse and Bernhardt, 2013). N₂O is produced via nitrification or denitrification depending on the availability of soil O₂ at the reaction site. N₂O can be further reduced to N₂ via denitrification under limited gaseous diffusion through water-filled pore space (Clough et al. 2005). Similar to CO₂ concentrations, changes in soil N₂O concentrations at 10 and 20 cm had an inverse linear relationship with simultaneous changes in soil O₂ at 10 and 20 cm. Greater soil O₂ depletion resulted in greater increases in N₂O concentrations at SD compared to QD (Figure 8e). Furthermore, the greatest increases in N₂O concentration occurred when soil O₂ depletion

exceeded 5%. Liptzen et al. (2011) also measured high N₂O concentrations at near-atmospheric soil O₂ concentrations in tropical soils. Thus, the higher N₂O concentrations found in our relatively oxic soil profile may be a result of nitrification or denitrification in anoxic microsites.

Consistently inundated soils that limit O₂ diffusion from the atmosphere can increase CH₄ production from methanogens and facilitate CH₄ transport via plants or ebullition (Whalen, 2005). Short-term saturation events, as those in this study, can stimulate both CH₄ production and oxidation. Thus, the CH₄ produced may be more readily oxidized during soil drying and lead to lower CH₄ flux from soil under varying hydrology compared to consistently inundated soil (Altor and Mitsch, 2008). Soil CH₄ concentrations can have a strong negative non-linear relationship with soil O₂ concentration when soil O₂ concentration is extremely low (<1%) for extended periods (months) in wet tropical soils (Liptzin et al., 2011). In our study, average soil O₂ depletion following precipitation ranged from 4.3 to 6.6% (Table 1), but we found changes in CH₄ concentration in the bulk soil were unaffected by short-term soil O₂ dynamics (Figure 8f). There are likely two non-mutually exclusive explanations for why CH₄ did not respond to decreased O₂: 1) the CH₄ produced in anoxic microsites was quickly oxidized before reaching the soil surface, and/or 2) the soil O₂ was not decreased significantly enough to deplete the redox buffering capacity of the soil to result in methanogenesis being the dominant form of anaerobic respiration. More frequent or longer periods of precipitation, extended water infiltration, or water table rise may lead to prolonged periods of low soil O₂, more robust methanogenic community, and greater soil CH₄ concentrations.

4.2. Using soil O₂ to understand biogeochemical hot spots and hot moments

Spatial and temporal variation in near-surface soil O₂ dynamics may be useful for understanding the presence (hot spots) and timing (hot moments) of disproportionately high

GHG surface fluxes. The spatial variability in soil O₂ depletion that we observed between and within SD and QD may be strongly linked to soil drainage patterns influenced by site-specific soil physical properties (i.e., soil texture) and topography. In addition to soil hydrologic properties, the spatial differences in soil O₂ depletion may reflect heterogeneity in microbial O₂ demand (Rubol et al., 2013). An increase in microbial O₂ demand under prolonged soil saturation would support our observations of greater increases in CO₂ and N₂O concentrations at SD during greater soil O₂ depletion. Furthermore, SD soil had significantly higher substrate induced respiration rates at 0-10 cm compared to QD (Table 1). Thus, our results support that both hydrologic properties and microbial O₂ demand are important drivers of subsurface GHG concentrations (i.e., CO₂ and N₂O).

While the spatial variability in soil O₂ depletion may provide new understanding of hot spots in GHG fluxes, temporal variability in soil O₂ dynamics may be more important for determining hot moments in GHG fluxes. Short-term soil O₂ fluctuation was clearly linked to subsurface CO₂ and N₂O concentration dynamics. The simultaneous decrease in CO₂ and N₂O concentrations that we observed with big gulps (Figure 8d,e) is not surprising because soil gases are diffusion limited during saturation and soil drainage lifts this limitation. When diffusion constraints were relaxed, we also observed an increase in surface CO₂ flux that co-occurs with big gulps (Figure 8a). This result supports that the timing of big gulps may play an important role in increased soil-atmospheric gas exchange (hot moments). Change in surface N₂O and CH₄ fluxes were not significantly related to short-term soil O₂ dynamics perhaps due to the weaker subsurface response of N₂O concentration and no response of CH₄ concentration during dynamic soil O₂ (Figure 8e,f). Since soils were never completely anaerobic in our study, the availability of O₂ may have limited the production of N₂O and CH₄ under short-term saturation.

The timing and magnitude of big gulps are associated with saturation duration and associated soil pore drainage driven by differences in soil properties such as soil texture and bulk density (e.g., QD vs. SD, Table 1). Furthermore, big gulps occur at a relatively consistent value of soil moisture near field capacity when soils begin to drain(Figure 3). This suggests that hot moments may depend on site-specific soil moisture patterns that correlate to big gulps. Others predict peak N₂O fluxes when soils are near field capacity due to the abundance of anaerobic microsites and increased diffusivity in the bulk soil (Davidson et al., 2000, McNicol and Silver, 2014). The field capacity at which max N₂O flux occurs can depend on differences in soil properties that influence soil drainage patterns (i.e., total porosity and bulk density) (Castellano et al., 2010). Therefore, determining the soil physical properties that influence the magnitude and timing of big gulps may also reveal the abiotic factors that control the magnitude and timing of hot moments in GHG emissions.

Surface GHG flux is a balance between the biogeochemical production/consumption of GHG and the transport of soil gases (Blagodatsky and Smith, 2012). Other studies also describe increased soil gas flux rates due to soil re-wetting when low GHG surface flux from dry soils precede rewetting events (e.g., Kim et al., 2012). Yet, scientists lack mechanistic understanding of the environmental drivers of the microbial-mediated process rates and position of reactions sites (Kuzyakov and Blagodatskaya, 2015), in addition to the physical transport of gases in the soil profile. With technical advancement and increased wide-spread use of soil sensor networks, soil O₂ dynamics show greater potential for gleaned new information on coupled biogeochemical and soil physical processes under varying soil hydrology.

Soil O₂ dynamics are reported across multiple ecosystems (Burgin and Groffman, 2012; Silver et al., 2012); yet, drivers of soil O₂ fluctuation are not well-represented in current

ecosystem models that simulate soil O₂ directly from soil moisture (Rubol et al., 2013). Unlike previous studies that measured soil O₂, we describe how rapid change in soil O₂ (big gulps) correspond to change in soil GHG fluxes. According to our study, soil O₂ dynamics can improve our mechanistic understanding of biogeochemical and physical shifts in the soil environment that influence hot spots and hot moments in GHG emission. The tight link between subsurface soil O₂ and CO₂ concentrations suggest that accurately modeling soil O₂ dynamics may further our capacity to predict rates of biogeochemical processes such as organic matter decomposition (Davidson et al., 2012). Additionally, near-continuous measurement of surface GHG fluxes may discern the temporal dynamics between subsurface GHG accumulation under soil O₂ depletion and short-term, enhanced rates of surface GHG fluxes under rapid soil gas transport. We recommend future efforts to collect soil GHG flux at higher temporal resolution following prolonged saturation (> 1 week) in order to highlight the influence of soil O₂ dynamics on biogeochemical hot moments.

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502 **Table Caption**

503 **Table 1.** Average $\pm$ standard error for soil physical properties and substrate induced respiration
504 at the Quick Draining (QD; n = 30) and Slow Draining (SD; n = 18) sites; *p<0.01 for QD and
505 SD t test at 0-10 cm; ** p<0.01 for QD and SD t test at 0-10 cm and 10-20 cm. Average $\pm$
506 standard error for soil O₂ depletion during saturation events, the duration of saturation events,
507 and soil temperature during saturation at QD 10 cm (n = 6 in 2013 and n = 20 in 2014), QD 20
508 cm (n = 4 in 2013 and n = 14 in 2014), SD 10 cm (n = 39 in 2014), and SD 20 cm (n = 29 in
509 2014).

Figure Captions

Figure 1. Distribution of percent sand at 0-10 cm from 120 soil cores collected across the post-construction wetland in 2011; site QD and SD were established in October 2013 and June 2014 and represent differences in soil texture across the wetland.

Figure 2. The design of subsurface gas sampling chambers, soil sensors, and surface flux chamber for each soil pit at QD and SD.

Figure 3. Soil O₂ and soil moisture from June - August 2014 at SD “r1” (a) 10 cm and (b) 20 cm.

Figure 4. Soil O₂ depletion following June-August 2014 precipitation events at (a) QD 10 cm, (b) SD 10 cm, (c) QD 20 cm, and (d) SD 20 cm.

Figure 5. Soil oxygen depletion as a function of mean temperature (°C) x saturation duration (days) for: (a) QD 10 cm, (b) SD 10 cm, (c) QD 20 cm, and (d) SD 20 cm following precipitation events in October 2013 and June-August 2014; *p<0.05.

Figure 6. Subsurface CO₂, N₂O, and CH₄, and O₂ concentration at (a) QD “r1” 10 cm, (b) SD “r1” 10 cm, (c) QD “r1” 20 cm, and (d) SD “r1” 20 cm.

Figure 7. Surface CO₂, N₂O, and CH₄ fluxes and O₂ concentration at 10 cm at (a) QD “r1” and (b) SD “r1”.

Figure 8. Change in surface (a) CO₂, (b) N₂O, and (c) CH₄ (c) fluxes and (d) subsurface CO₂, (e) N₂O, and (f) CH₄ concentrations (10 and 20 cm) as a function of change in soil O₂ concentration (*p<=0.05) during October 2013 and from June-August 2014.

530 **Table**

	QD 0-10 cm	QD 10-20 cm	SD 0-10 cm	SD 10-20 cm
Bulk Density ** (g cm ⁻³)	1.26 ± 0.05	1.78 ± 0.03	1.40 ± 0.03	1.63 ± 0.02
% Sand **	52.76 ± 3.61	45.65 ± 4.83	22.61 ± 0.57	30.72 ± 3.19
% Silt *	24.36 ± 3.52	34.00 ± 4.62	52.96 ± 0.56	39.20 ± 3.39
% Clay **	22.87 ± 0.45	20.34 ± 0.69	24.42 ± 0.54	30.09 ± 0.87
% Carbon	2.21 ± 0.12	1.61 ± 0.09	2.18 ± 0.07	1.66 ± 0.02
% Nitrogen **	0.13 ± 0.006	0.13 ± 0.003	0.21 ± 0.004	0.17 ± 0.002
Substrate Induced Respiration * (CO ₂ mg C g ⁻¹ soil hr ⁻¹)	0.05 ± 0.01	0.04 ± 0.004	0.25 ± 0.01	0.06 ± 0.01
	QD 10 cm	QD 20 cm	SD 10 cm	SD 20 cm
O ₂ depletion (%)	4.29 ± 1.57 (2013) 4.55 ± 0.46 (2014)	4.57 ± 1.02 (2013) 4.36 ± 0.57 (2014)	5.77 ± 0.93 (2014)	6.59 ± 1.13 (2014)
Saturation duration (days)	0.8 ± 0.13 (2014) 1.77 ± 11.65 (2013)	0.83 ± 0.17 (2014) 4.11 ± 1.17 (2013)	1.04 ± 0.13 (2014)	2.54 ± 0.46 (2014)
Soil temperature (°C)	22.23 ± 0.28 (2014) 15.47 ± 1.63 (2013)	22.09 ± 0.34 (2014) 12.17 ± 1.69 (2013)	23.40 ± 0.19 (2014)	22.86 ± 0.14 (2014)

531 **Table 1.**

Figure

Figure 2

= soil gas chamber (CO_2 , CH_4 , and N_2O)

= hydra probe (soil moisture)

= Apogee sensor (soil O_2)

Figure 3

Figure 4

Figure 5

Figure 6

Figure 8