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Spatial and temporal variability in soil CO_2–C emissions and relation to soil temperature at King George Island, maritime Antarctica

Newton La Scala Junior [a,*], Eduardo de Sá Mendonça [b], Juliana Vanir de Souza [c], Alan Rodrigo Panosso [a], Felipe N.B. Simas [c], Carlos E.G.R. Schaefer [c]

[a] *FCAV/UNESP, Via de Acesso, Prof. Paulo Donato Castellane s/n 14884-900 Jaboticabal, SP, Brazil*
[b] *Department of Plant Production, Federal University of Espírito Santo, 29500-000 Alegre, ES, Brazil*
[c] *Soil Science Department, Federal University of Viçosa, Av. PH Rolfs, sn. 36570-000 Viçosa, Minas Gerais, Brazil*

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Abstract

Few studies have examined the effects of temperature on spatial and temporal trends in soil CO_2–C emissions in Antarctica. In this work, we present *in situ* measurements of CO_2–C emissions and assess their relation with soil temperature, using dynamic chambers. We found an exponential relation between CO_2 emissions and soil temperature, with the value of Q_{10} being close to 2.1. Mean emission rates were as low as 0.026 and 0.072 g of CO_2–C m^{-2} h^{-1} for bare soil and soil covered with moss, respectively, and as high as 0.162 g of CO_2–C m^{-2} h^{-1} for soil covered with grass, *Deschampsia antarctica* Desv. (Poaceae). A spatial variability analysis conducted using a 60-point grid, for an area with mosses (*Sannionia uncianata)* and *D. antarctica*, yielded a spherical semivariogram model for CO_2–C emissions with a range of 1 m. The results suggest that soil temperature is a controlling factor on temporal variations in soil CO_2–C emissions, although spatial variations appear to be more strongly related to the distribution of vegetation types.

Keywords: Climate change; Cryosols; Gelisols; Soil carbon stock; Soil respiration

1. Introduction

Soil CO_2 emissions are a major contributor to the global carbon cycle, as up to 100 Pg C yr^{-1} is emitted by CO_2 (Raich and Schlesinger, 1992). Small changes in soil CO_2 emissions could have a strong influence on increases in CO_2 atmospheric, resulting in an enhanced greenhouse effect. In temperate and polar regions, soil temperature is the main controlling factor on temporal and spatial variability in soil CO_2 emissions (Reichstein et al., 2003). The response of soil CO_2 flux to temperature has been shown to vary, both spatially and temporally, in several ecosystems (Buchmann, 2000; Kirschbaum, 1995; Rustad et al., 2000; Schleser, 1982; Trumbore et al., 1996). Despite the important influence of soil temperature on temporal variations in soil CO_2 emissions, the nature of this

* Corresponding author.
E-mail address: lascala@fcav.unesp.br (N. La Scala Junior).

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relation is uncertain in the Antarctic region, as few studies have investigated the quality, quantity, and dynamics of organic matter in Antarctic soils (Beyer et al., 2000; Blume et al., 2001; Hopkins et al., 2009). Such studies are important because maritime Antarctica has the highest mean temperature and precipitation of any region in Antarctica, resulting in relatively high primary productivity, enhanced pedogenesis, and high biological activity in general.

Soils affected by bird activity, the so-called ornithogenic Cryosols, constitute the main organic C reservoirs in Antarctic terrestrial ecosystems (Michel et al., 2006; Simas et al., 2006, 2007a; Tatur et al., 1997). However, few studies have assessed the carbon stocks of permafrost in maritime Antarctica, although recent works have shown that the accumulation rate of organic matter in soils from coastal Antarctica is higher than previously expected (e.g., Simas et al., 2007b). A large proportion of the C stored in Cryosols is protected in permafrost (Michaelson et al., 2004), suggesting a high potential for CO_2−C emissions in face of the current global warming scenario, which could result in the degradation of permafrost. Laboratory experiments indicate a ten-fold increase in the respiratory rate for Arctic Cryosols after an increase in soil temperature from −0.5 to 0 °C, a four-fold increase with an increase in temperature from 0 to 5 °C, and a two-fold increase with an increase from 0 to 25 °C (Michaelson et al., 2004). Previous results have shown that soil temperature governs soil carbon losses for most of the summer period, when soils are exposed and not covered by snow (Hopkins et al., 2006; Park and Day, 2007).

The decomposition of soil organic matter has been modeled based on variations in soil aeration, the quality of organic matter, the degree of physical protection of labile carbon, soil water content, and temperature (Parton et al., 1987). However, under natural conditions, without human influence, the loss of soil C depends almost exclusively on soil temperature and water content. Therefore, the sensitivity of soil C to soil temperature is affected by numerous factors that are directly or indirectly related to temperature (Yuste et al., 2007). This sensitivity is commonly quantified in terms of the Q_{10} factor, which represents the change in emissions due to a temperature increase of 10 °C, and is even more commonly expressed by exponential models of the relation between these variables (Fang and Moncrieff, 2001; Xu and Qi, 2001; Lloyd and Taylor, 1994). Despite the efforts of previous works, most studies concerned with soil CO_2−C fluxes in Antarctica have been based on incubation experiments, with temperatures controlled under laboratory conditions. It is also important to perform *in situ* studies in Antarctica, especially when conducted intensively in time and space, as this region has recorded the highest temperature increases worldwide in recent decades (Turner et al., 2007).

Geostatistical analyses have been successfully applied to characterization of the spatial variability structure of several physical, chemical, and biological properties of soils (Johnson et al., 1996; Sinegani et al., 2005; Wang et al., 2002), including CO_2 emissions (Elberling, 2007). The derivation of a function that relates the covariance of the studied property with the distance between points in the spatial variability analysis has the advantage of enabling more accurate estimations of the studied property (Webster and Oliver, 1990). Understanding the spatial variability of soil CO_2−C emissions is important to gain a better understanding of the dynamics of CO_2−C in various ecosystems, and the characterization of spatial variability assists in interpreting such phenomena at a given scale.

Soil organic matter and porosity are also thought to influence the spatial variability of soil CO_2−C emissions (Fang et al., 1998; La Scala et al., 2000; Schwendenmann et al., 2003; Xu and Qi, 2001). However, in vegetated environments where soil temperature is a controlling factor on emissions, soil temperature could also influence the spatial distribution of soil organic matter and consequently soil CO_2−C emissions. Few studies have sought to characterize the spatial variability structure of soil CO_2 emissions using semivariance, especially in the Antarctic environment.

In the present work, we performed *in situ* analyses of temporal and spatial trends in soil CO_2−C emissions in maritime Antarctica, and examined their relation to temperature variations during the austral summer of 2008/2009.

2. Materials and methods

2.1. Site description

The studied site was selected based on previous soil studies conducted in ice-free areas of Admiralty Bay, King George Island (Michel et al., 2006; Simas et al., 2006, 2007a; see also Fig. 1a). The studied soil is a Leptic Thimorphic Cryosol (according to the WRB classification scheme; see also Fig. 1b) located at 62°04′S, 58°24′W, in the area surrounding the Comandante Ferraz Brazilian Station, located on that part of the Keller Peninsula consisting of sulfide-

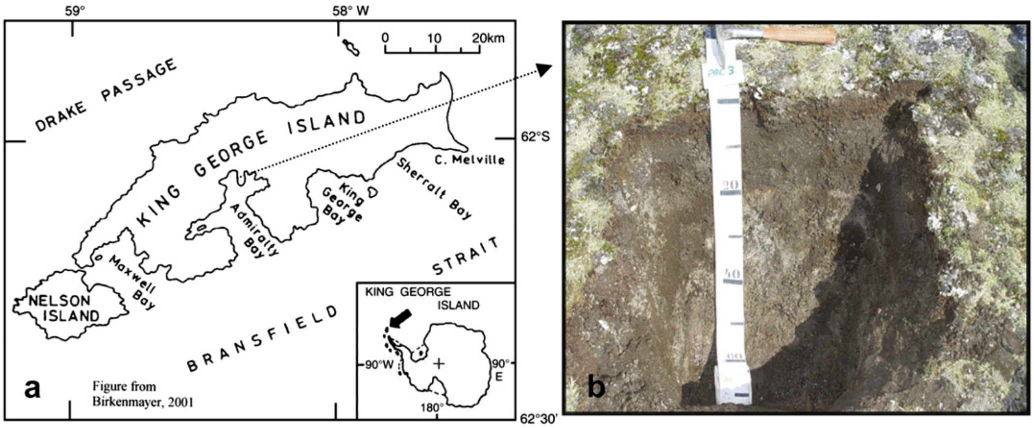

Fig. 1. (a) The Admiralty Bay, King George Island and (b) Leptic Timorphic Cryosol soil layer.

bearing andesites. Datasets from 1982 to 2002, acquired at the station, yield mean daily maximum air temperatures varying from −6.4 °C in July to +2.3 °C in February, with mean annual precipitation of 366.7 mm. Soils in this area are acidic and highly weathered by Antarctic standards. The temperature remains above freezing throughout summer, meaning that plant communities are able to become established and grow during this period (Simas et al., 2007a). The study site is covered by patches of moss carpet dominated by *Sannionia uncianata*, grass carpets of *Deschampsia antarctica* Desv. (Poaceae), and bare soil. A typical soil profile is described and samples were collected at depths of 0–10 and 10–20 cm for analyses of total organic C.

2.2. In situ measurement of soil CO_2–C fluxes

2.2.1. Temporal variability

Measurements were made during the Antarctic summer, from 8th January to 15th March 2008, which is the period of maximum soil exposure and highest biological activity in maritime Antarctica. Nine PVC soil collars (diameter = 10 cm) were placed in the soil to a depth of approximately 3 cm, representing patches colonized by *D. antarctica*, moss carpet, and bare soil. At these sites, CO_2–C fluxes were measured using a portable LI-8100 analyzer (LiCor, Lincoln, NE/USA) coupled to a dynamic chamber (LI-8100-102), collectively constituting a survey chamber. Measurements of emissions at each survey point were based on three replications of a single measurement over 1.5 min, during which time measurements were made of CO_2 concentrations inside the chamber at 3-s intervals. An interpolation procedure was performed to estimate the emission rate at each point. The soil chamber, which has an internal volume of 854.2 cm^3 and a circular contact area with the soil (83.7 cm^2), was placed on PVC soil collars previously inserted at each point. Measurements were made on 20 days during the study period, when soil temperature varied between approximately −1 and 8 °C. Soil temperature in the 0–10 cm layer was also measured at all study points using a pen-like digital thermometer (DELLT, model DT 625, Brazil).

2.2.2. Spatial variability

Spatial variability in soil CO_2–C emissions was analyzed from 09:40 to 11:00 on 13 March 2009. Measurements were made with a dynamic chamber of 3 × 1.5 m in size, using a 60-point regular grid with a minimum distance of 0.5 m between grid points. The grid was installed in such a way that 30 of the points were located on moss carpet and 30 on patches of *Deschampsia* grass. In several cases, collars (10 cm in diameter) were placed in areas with suppressed vegetation or fewer tuffs, because of the natural heterogeneity of the site. The aboveground parts of plants were not removed, meaning that the data include aboveground respiration in some cases.

Spatial variability was analyzed using descriptive statistics and by adjusting semivariogram models to data on soil CO_2–C emissions and soil temperature. The semivariance (Isaaks and Srivastava, 1989) was calculated as follows:

$$\gamma(h) = \frac{1}{2N(h)} \sum_{i=1}^{N(h)} [z(x_i) - z(x_i + h)]^2 \qquad (1)$$

where N is the number of pairs, separated by distance h, between sample points in the soil; $\gamma(h)$is the semivariance at a separation distance h; $N(h)$ is the number of pairs of points separated by h; $z(x_i)$ is the property value at point x_i; and $z(x_i + h)$ is the property value at point $x_i + h$.

A semivariogram graph may have a random or a systematic behavior that can be described by theoretical models (e.g., spherical, Gaussian, exponential). Experimental semivariograms were created from calculated values of $\gamma(h)$ for all pairs of points separated by distance h, which were adjusted by the following mathematical models:

(a) exponential:

$$\gamma(h) = C_0 + C\{1 - \exp[-3(h/a)]\},\ h > 0 \tag{2.1}$$

(b) spherical:

$$\begin{aligned}\gamma(h) &= C_0 + C\left[3/2(h/a) - 1/2(h/a)^3\right], \quad 0 \le h \le a \\ \gamma(h) &= C_0 + C, \quad h > a\end{aligned} \tag{2.2}$$

(c) Gaussian:

$$\gamma(h) = C_0 + C\left\{1 - \exp\left[-3(h/a)^2\right]\right\}, 0 < h < d \tag{2.3}$$

where d is the maximum distance over which the semivariogram is defined and a is the so-called range distance (Isaaks and Srivastava, 1989). In this case, the semivariance value increases with increasing separation of the points (distance h) until a distance is reached at which the sill, that is the semivariance value in which the semivariogram curve stabilizes, and (C_0 + C_1) remains constant. The distance at which this stabilization occurs is called the range distance (a). The pure nugget effect (C_0) is the value at which the adjusted theoretical model crosses the y axis.

The model adjusted to the semivariogram was used to generate the so-called kriging map by interpolation, thereby providing estimates of the studied property at non-sampled sites. This process is related to estimations based on the property values of closest neighbors, using adjusted theoretical semivariogram models (Webster and Oliver, 1990).

2.3. Determination of soil total organic carbon

Total organic carbon (TOC) was quantified for the 0–10 and 10–20 cm layers. Samples were air-dried, ground, and passed through a 0.20 mm sieve. TOC was quantified by wet combustion with an external heat source (Yeomans and Bremner, 1988). Carbon stocks were calculated by the following relation: Cst = (TOC) × (D) × (L) × (1 − % fragments > 2 mm/100), where Cst is the C stock (kg m^{-2}), TOC is expressed in gC g soil^{-1}, D is soil density (kg m^{-3}), and L is layer thickness (m).

The density of the soil fraction that passed through the 2 mm sieve was measured in laboratory. TOC content was 7.57 and 5.96 g kg^{-1} in the 0–10 and 10–20 cm layers, respectively. The total C stored in the upper and lower layers was estimated to be 505 and 245 g C m^{-2}, respectively, representing the mean stored C for the study site.

2.4. Statistical analyses

Descriptive statistics of CO_2–C emissions, as well as graphs and regression equations, were obtained using Origin 6.0$^{@}$ software (MicroCal). Spatial variability models were derived using GS + software (Gamma Design, 1998) and kriging maps were obtained with Surfer software (GOLDEN SOFTWARE, 1995). The studied properties were submitted to one-way analysis of variance and mean separation procedure (SAS Institute, 1991).

3. Results and discussion

3.1. Temporal variability

The mean soil CO_2–C emissions were 0.162 and 0.026 g of CO_2–C m^{-2} h^{-1} (3.74 and 0.61 μmol CO_2 m^{-2} s^{-1}) for sites with *D. antarctica* and bare soil, respectively (Table 1). For the moss carpet, an intermediate value of 0.072 g of CO_2–C m^{-2} h^{-1} (1.67 μmol CO_2 m^{-2} s^{-1}) was obtained. The three values are significantly different from each other ($p < 0.05$), suggesting that CO_2–C emissions are determined by the type of plant community at a given site.

Sites with *Deschampsia* and those with mosses shows greater temporal variability and yielded CO_2 emissions that were 514 and 174% greater than those at bare sites, respectively. This finding reflects the effect of plant respiration on CO_2–C emissions. The finding of the highest value for the *Deschampsia* site is partly because it is a higher plant and therefore possesses a true root system, which increases respiration. Soil thawing, longer daylight hours, and higher temperatures during the austral summer favor biological activity and the liberation of easily decomposable root exudates by *D. antarctica*.

A significant exponential relation ($p < 0.05$) was obtained between CO_2–C emissions and temperature for all three studied surface covers (Fig. 2 and Table 2), yielding a significant relation with emissions. Therefore, a regression was performed using $FCO_2 - C =$

Table 1
Descriptive statistics of CO_2–C (g m^{-2} h^{-1}) emissions for the studied soil covers.

	Mean	Standard deviation	Standard error	Min.	Max.
D	0.162	0.049	0.011	0.073	0.255
M	0.072	0.027	0.006	0.037	0.124
B	0.026	0.009	0.002	0.015	0.045

D = *Deschampsia antarctica*; M = mosses; B = bare soil. $n = 22$.

$F_0 e^{A2 \times T_{soil}}$, where F_0 is the initial emission (at $T_{soil} = 0$) and A2 is the sensitivity of emissions to soil temperature (°C^{-1}). After linearization, the equation becomes Ln(FCO_2–C) = A1 + A2(T_{soil}), where A1, the linear coefficient, is equal to Ln(F_0).

The A2 coefficients were statistically similar for all of the studied situations. Therefore, it was possible to estimate a mean value of 0.073 °C^{-1} for A2. Based on this value, an increase in soil CO_2–C emissions of 7.6% was predicted for each 1 °C increase in soil temperature, when close to the melting point in this part of Antarctica. By comparison, the global warming scenario indicates a 3.1 °C increase in air temperature by the end of the 21st century (Chapman and Walshj, 2006).

Q_{10} was calculated as

$$Q_{10} = e^{10 \times A2} \quad (3)$$

The highest Q_{10} value (Table 2) was obtained for bare soil (2.21), and the lowest for the area colonized with *Deschampsia* (1.98). Because the values of A2 ± standard deviation for the different situations are overlapping values, the Q_{10} values are not significantly different among the stations. This result differs from that reported by Boone et al. (1998), who found that Q_{10} values in a temperate forest varied with the contribution of root respiration.

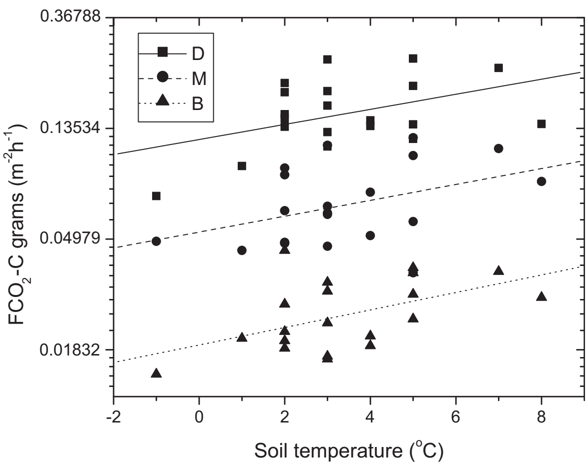

Fig. 2. CO_2–C emissions plotted against soil temperature for the studied soil covers. The straight lines are linear regression adjustments for D = *Deschampsia antarctica*, M = mosses, and B = bare soil. $n = 22$.

The occurrence of photosynthetic activity would make it difficult to determine the relation between CO_2–C emissions and soil temperature, especially at vegetated sites. An exponential regression yielded lower determination coefficients than those reported in similar works on Antarctic soils (Hopkins et al., 2006; Smith, 2005), possibly because our determinations were made *in situ,* whereas most previous studies were performed in a laboratory.

The Q_{10} values of the present study are similar to those reported previously. For example, in a laboratory experiment, Hopkins et al. (2006) reported Q_{10} values ranging from 1.4 to 3.3 for different Antarctic soils submitted to temperature increases ranging from −0.5 to 20 °C. In a study of various soils from maritime Antarctica, Smith (2005) reported that the emission rate increased exponentially with increasing soil temperature, regardless of water content, although water content had a strong influence on Q_{10}, which varied from 1.59 to 2.51. In the present work, the sensitivity of CO_2–C emissions to soil temperature is calculated for a soil condition without the influence of water content, as we consider ice-free areas during the thawing period, for which Q_{10} is close to 2, similar to the value reported by Yuste et al. (2007).

The observed increase in CO_2–C emissions with increasing temperature indicates an increase in microbiological activity, which enhances the mineralization of soil organic matter (Silva and Mendonça, 2007). This is only part of the potential global warming scenario, as increases in the rate of photosynthesis and net primary production may occur, thereby increasing the input of organic matter into the soil (Michel et al., 2006). Therefore, the final soil TOC content and the net balance between C sequestration and emission would depend on the equilibrium attained between the input of organic matter and the mineralization rate under higher-temperature conditions.

3.2. Spatial variability models of CO_2–C emissions and soil temperature

Table 3 lists the descriptive statistics and the results of spatial variability analysis of soil CO_2–C emissions and soil temperature based on grid measurements. The observed mean emission value was 0.061 g CO_2–C m^{-2} h^{-1} (1.41 µmol CO_2 m^{-2} s^{-1}) and the mean temperature was 4.94 °C. The mean emission rate is slightly higher than that observed in

Table 2
Parameters (±standard error) of the model between CO_2–C emissions and soil temperature for the studied soil covers.

Soil Cover	$\ln(FCO_2-C) = A1 + A2\ T_{soil}$		R	P	Q_{10}
	A1	A2			
D	−2.10013 ± 0.127	0.06806 ± 0.03215	0.45	0.048	1.98 ± 0.65
M	−2.93624 ± 0.14957	0.07164 ± 0.03787	0.41	0.075	2.05 ± 0.80
B	−3.95657 ± 0.12399	0.07913 ± 0.03139	0.51	0.021	2.21 ± 0.71

$n = 22$; [A1] = adimensional; [A2] = $°C^{-1}$; R = linear correlation coefficient; P = significance coefficient; D = *Deschampsia antarctica*; M = mosses; B = bare soil.

the temporal variability study, conducted at the same site in the previous year, when considering the measurements made at temperatures close to 5 °C. The results also show that the coefficient of variation (CV) for soil CO_2–C emissions is higher than that for soil temperature (Table 3). This result has been commonly observed in similar studies that simultaneously analyzed the spatial distribution of both properties (Epron et al., 2004; Konda et al., 2008; La Scala et al., 2000). The points at which low emissions were recorded were generally those under less-dense vegetation inside the collar chamber, emphasizing the direct effect of vegetation on CO_2 emissions, due mainly to respiration by plant roots, as observed by Welker et al. (2000) in the Arctic tundra. According to the spatial variability criteria defined by Warrick and Nielsen (1980), the CV values obtained for CO_2–C emissions can be considered high, as they exceed 24%. This value is in agreement with those found in other parts of the world, under various vegetation covers (from forest to bare soils) (Epron et al., 2004; Konda et al., 2008; La Scala et al., 2000; Panosso et al., 2008; Schwendenmann et al., 2003; Tedeschi et al., 2006). Similarly, the CV value obtained for soil temperature can be considered high (CV > 24%).

The adjusted semivariogram parameters indicate that the models were spherical and Gaussian for soil CO_2–C emissions and soil temperature, respectively, with both adjustments yielding high determination coefficients ($R^2 > 0.95$). Most of the spatial variability models adjusted to soil CO_2 emissions are spherical or exponential (Ishizuka et al., 2005; Konda et al., 2008; Kosugi et al., 2007; La Scala et al., 2000; Stoyan et al., 2000; Tedeschi et al., 2006).

Isaaks and Srivastava (1989) reported that exponential models adjust better to more erratic phenomena, while Gaussian and spherical models perform better in describing phenomena with greater continuity, without large changes at small scales. The scale dependence degree (SDD) of the data (Table 3), calculated as $C_0/C_0 + C_1$, indicates the moderate spatial dependence (Cambardella et al., 1994) of soil CO_2–C emissions (0.25 < SDD < 0.75) but strong spatial dependence of soil temperature (SDD < 0.25). Moderate and weak SDD values have been observed previously for soil CO_2–C emissions in various ecosystems (Ishizuka et al., 2005; Konda et al., 2008; Kosugi et al., 2007; La Scala et al., 2000; Stoyan et al., 2000).

Range values were 1.00 and 3.03 m for CO_2–C emissions and soil temperature, respectively, indicating that soil temperature at the site shows little relation with soil CO_2–C emissions, since the models and ranges are not similar. In addition, no significant linear correlation is found between the 60-point measurements of CO_2–C emissions and soil temperature. This finding indicates that the soil CO_2–C emission variability model cannot be directly related to spatial changes in soil temperature, especially under vegetated conditions. Therefore, although soil temperature is a controlling factor regarding temporal variability in soil CO_2–C emissions, it shows no relation to spatial

Table 3
Mean ± standard error (g CO_2–C $m^{-2}\ h^{-1}$), coefficient of variation (CV, %), models, and estimated parameters of derived soil CO_2–C emissions and soil temperature semivariograms.

Mean ± SE	CV	Model	C_0	$C_0 + C_1$	a (m)	SSR	r^2	SDD
Soil CO_2–C emissions								
0.061 ± 0.004	41	Spherical	0.003	0.011	1.00	1.79 E − 07	0.99	0.27
Soil Temperature								
4.94 ± 0.16	25	Gaussian	0.1840	2.3620	3.03	2.35 E − 01	0.96	0.08

$n = 60$; SSR: sum of the square of residues. SDD: spatial dependence degree = $C_0/(C_0 + C_1)$. We used the following classification scheme: "strong" for values below 0.25, "moderate" for values between 0.25 and 0.75, and "weak" for values above 0.75 (Cambardella et al., 1994).

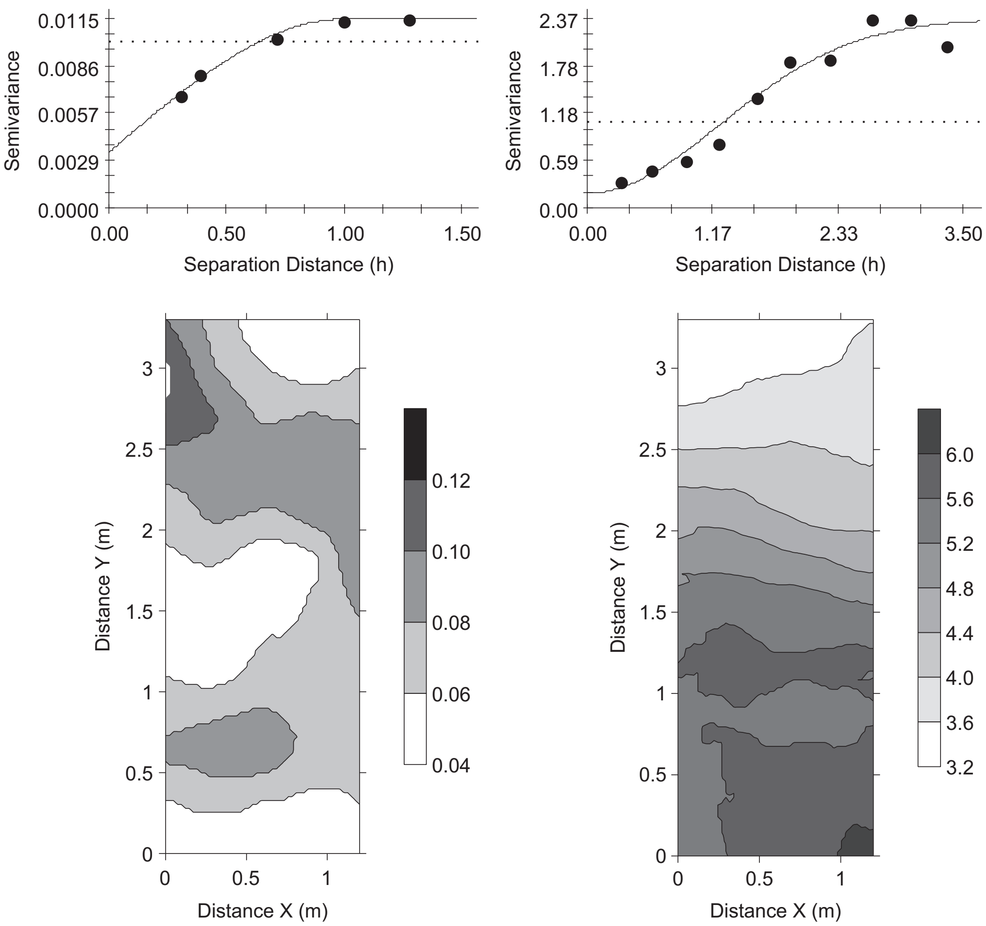

Fig. 3. Semivariance as a function of separation distance (upper) and kriging maps of soil CO_2–C emissions (lower left, g m^{-2} h^{-1}) and soil temperature (lower right, °C).

variability in soil CO_2–C emissions in the studied environments, at least under vegetated conditions.

Fig. 3 shows semivariograms and kriging maps of CO_2–C emissions and soil temperature. The 30 points under *D. antarctica* yield higher mean emission rates (0.064 g CO_2–C m^{-2} h^{-1} or 1.49 μmol CO_2 m^{-2} s^{-1}) than do the 30 points under mosses (0.057 g CO_2–C m^{-2} h^{-1} or 1.32 μmol CO_2 m^{-2} s^{-1}). This feature is also observable in the CO_2–C emission map (Fig. 3), which shows higher emissions in its upper part, where *D. antarctica* was concentrated. A comparison of the two maps in Fig. 3 reveals that higher soil temperatures were generally recorded at the area under mosses than under *D. antarctica*. Nevertheless, this does not appear to have strongly influenced CO_2 emissions, suggesting that the difference in emission rate between the soil covers is derived mainly from root respiration rather than soil carbon decay.

4. Conclusions

In a maritime Antarctica soil, we found the highest mean CO_2–C emission rate at a site with *D. antarctica*, compared with sites with moss carpet and bare soil, indicating the effect of soil respiration by the root system.

We found an exponential relation between CO_2 emissions and soil temperature. The highest sensitivity of emissions to soil temperature was found for bare soil, indicating the emission of C present in the soil, independent of vegetation.

The results of a spatial variability analysis suggest that the type of vegetation, rather than soil temperature, controls the spatial variability model, because no relation was observed between soil CO_2–C emission and soil temperature, either by linear correlation or by comparing the spatial variability models and maps.

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