Soil moisture influenced the interannual variation in temperature sensitivity of soil organic carbon mineralization in the Loess Plateau

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Abstract

Temperature sensitivity of soil organic carbon (SOC) mineralization (i.e., Q_{10}) determines how strong the feedback from global warming may be on the atmospheric CO_2 concentration, thus understanding the factors influencing the interannual variation in Q_{10} is important to accurately estimate local soil carbon cycle. *In situ* SOC mineralization rate was measured using an automated CO_2 flux system (Li-8100) in long-term bare fallow soil in the Loess Plateau (35°12′ N, 107°40′ E) in Changwu, Shaanxi, China from 2008 to 2013. The results showed that the annual cumulative SOC mineralization ranged from 226 to 298 g C m^{-2} y^{-1}, with a mean of 253 g C m^{-2} y^{-1} and a CV of 13%, annual Q_{10} ranged from 1.48 to 1.94, with a mean of 1.70 and a CV of 10%, and annual soil moisture content ranged from 38.6 to 50.7% soil water-filled pore space (WFPS), with a mean of 43.8% WFPS and a CV of 11%, which were mainly affected by the frequency and distribution of precipitation. Annual Q_{10} showed a quadratic correlation with annual mean soil moisture content. In conclusion, understanding of the relationships between interannual variation in Q_{10}, soil moisture and precipitation are important to accurately estimate the local carbon cycle, especially under the changing climate.

Keywords: Soil temperature; SOC mineralization; distribution and frequency of precipitation.

1. Introduction

Temperature sensitivity of soil organic carbon (SOC) mineralization (hereafter refer to as Q_{10}) is of critical importance because it determines how strong the feedback from global warming may be on the atmospheric CO_2 concentration (Ågren and Wetterstedt, 2007). However, this is an issue of considerable debatable (Davidson et al., 2006; Kirschbaum, 2006), and the variations in Q_{10} are the main source of controversies in this feedback intensity (Larionova et al., 2007; Karhu et al., 2010; Conant et al., 2011; Sakurai et al., 2012). Therefore, understanding the factors influencing Q_{10} is important to accurately estimate C cycle and the feedback from the expected warmer climate.

Previous studies have shown that Q_{10} variations are closely related to soil temperature (Kirschbaum, 2006; Von Lutzow and Kogel-Knabner, 2009), substrate availability (Ågren and Wetterstedt, 2007; Gershenson et al., 2009), substrate quality (Von Lutzow and Kogel-Knabner, 2009; Sakurai et al., 2012), and the size and composition of microbial population (Djukic et al., 2010; Karhu et al., 2010). Soil moisture is the most significant limiting factor for underground physiological processes in dry and semi-dry ecosystems (Balogh et al., 2011; Cable et al., 2011; Wang et al., 2014). Soil water availability may indirectly affect Q_{10} by influencing the diffusion of substrates, because the diffusion of extracellular enzymes produced by microorganisms and available substrates must conduct in the liquid phase (Davidson et al., 1998; Illeris et al., 2004), but the response of Q_{10} to soil water availability is extremely complex and controversial (Davidson et al., 2000; Davidson et al., 2006; McCulley et al., 2007). For example, Gulledge and Schimel (2000) found that Q_{10} was larger in wet years than in drought years, whereas the opposite result was found by Dorr and Mdnich (1987). However, many other studies that mainly focused on the short-term or seasonal variation in Q_{10} (Davidson et al., 2006) have showed that Q_{10} was not affected by soil moisture (Fang and Moncrieff, 2001; Reichstein et al., 2002; Jassal et al., 2008). Additionally, soil water availability experienced marked seasonal and interannual fluctuations in these ecosystems due to uneven rainfall distribution caused by the abnormal increase of atmospheric CO_2 concentrations (Solomon et al., 2007). The uneven rainfall distribution inevitably influenced soil moisture availability (Coronato and Bertiller, 1996; Qiu et al., 2001; Cho and Choi, 2014). Xiao et al. (2014) have shown that the interannual changes in soil moisture storage in the Loess Plateau were decided by the difference in soil moisture storage between October and April, because precipitation from April to October of 2004 to 2010 accounted for at least 86% of annual rainfall. However, to our knowledge, there have been few studies investigating the relationship between interannual variation in Q_{10} and soil moisture under natural conditions.

The Loess Plateau is located in northwest China covering an area of 640,000 km^2. It has a continental monsoonal climate and shows a dramatically interannual fluctuations in precipitation, with the highest precipitation of 1262 mm and the lowest precipitation of only 80 mm, and a mean value of 150–750 mm (Lin and Wang, 2007). The precipitation in the loess regions also shows a dramatically seasonal variation, and approximately 60%–80% of the annual precipitation falls during the three summer months from July to September (Guo et al., 2012). Several recent studies have attempted to determine the dominant factors responsible for the variation of soil respiration in vegetation ecosystems (Lafond et al., 2011; Shi et al., 2011; Jurasinski et al., 2012). However, there have been no studies on the interannual variation in Q_{10}, nor the factors responsible for these changes. This highlights the need to accurately evaluate the response of SOC mineralization to increasing temperature under warmer climate scenarios in the eroded or degraded regions, because air temperature has been increasing over the past decades (Fan and Wang, 2011; Wang et al., 2012). Thus, the objectives of the present study are to (1) quantify the interannual variation in Q_{10}; (2) determine the effect of soil moisture on this interannual variation for the period 2008–2013 in the Loess Plateau, China.

2. Materials and methods

2.1 Site description

This study was a part of a long-term field experiment that began in 1984 in the State Key Agro-Ecological Experimental Station in the Loess Plateau in Changwu, Shaanxi, China (35°12′ N, 107°40′ E; 1,200 m above sea level) (Fig. 1). This region had a continental monsoon climate with a mean annual precipitation of 560 mm for the period 1984–2013, over 60% of which occurred from July to September. During this 30-year period, the annual mean air temperature was 9.4 ℃ and the monthly mean temperature between July and September was 19.4 ℃. The study site is also characterized by a ≥10 ℃ accumulated temperature of 3029 ℃, an annual sunshine duration of 2230 h, an annual total radiation of 484 kJ cm^{-2}, and a frost-free period of 171 days.

The site was located in a typical rain-fed cropping region of the Loess Plateau highland in northwest China. The soil was classified as a loam (Cumulic Haplustoll, USDA Soil Taxonomy System) developed from loess deposits. Soils collected at the study site in 1984 at a depth of 0–20 cm contained 10.5% $CaCO_3$, 6.5 g organic C kg^{-1}, 0.80 g total N kg^{-1}, and 200 mg NH_4OAc-extractable K kg^{-1}, 3.0 g kg^{-1} available phosphorus, and had a pH of 8.4 (with a 1: 1 ratio of soil: H_2O), a water-holding capacity of 0.29 cm^3 cm^{-3} (v/v), the wilting point of 11%, a soil bulk density of 1.3 g cm^{-3}, soil porosity of 51%, and a clay content of 24%.

2.2 Experimental design and management

This study was a part of a long-term field experiment established in June 1984. The plot used in the present study is taken from a bare plot in a state of fallow since June 1984 after the harvesting of winter wheat (*Triticum aestivum* L. ‘Chang Wu 131 series’), and living weed was artificially removed timely. Therefore, there were no vegetation or inputs of aboveground and belowground litter, and then SOC mineralization rates in the bare fallow soil did not include root respiration and litter mineralization and decomposition. In this paper, three bare fallow plots were used to investigate the mechanism of underground SOC mineralization rates. All plots of 10.3 m × 6.5 m (66.95 m^2) were randomly arranged in three blocks. The plots were separated by 0.5 m spaces, whereas the blocks were separated by 1 m strips.

2.3 Measurements of SOC mineralization rate and soil microclimate

SOC mineralization rate was measured using an automated closed soil CO_2 flux system with a portable chamber (20 cm in diameter, Li-8100, Lincoln, NE, USA). Approximately one day before the first measurement, a polyvinyl chloride (PVC) collar (20 cm in diameter and 12 cm in height) was inserted to a depth of 2 cm into each plot, and left in place throughout the experimental period from 2008 to 2013. Although previous studies have demonstrated a significant spatial variation of soil respiration, especially in the sites with complex terrain (causing the redistribution of SOC) and different vegetation types (Epron et al., 2006; Luan et al., 2012), the spatial variation of SOC mineralization rate in our sites is small with a variation coefficient of only 4% and 5% in summer and winter, respectively (Table 1). This can be attributed to that there have been no vegetation or inputs of (aboveground and belowground) litter in our plots since 1984 (absolute fallow), and the soil was derived aeolian deposit loess and flat terrain. Due to the small areas of our plots (66.95 m^2) and time constraints (5 min for measuring SOC mineralization rate in a given PVC collar), only one PVC collar was used in each plot for measuring SOC mineralization rate. All visible living organisms were removed before the measurement. If necessary, one or more additional measurements would be taken until the variations between two consecutive measurements were less than 15%. The final instantaneous soil respiration for a given collar was the average of the two measurements with a 90 s enclosure period and 30 s delay between them. Field measurements were performed between 09:00 and 11:00 AM from March 2008 to November 2013, except in December, January, and February because of cold weather. A total of 17, 25, 26, 22, 26 and 17 SOC mineralization measurements were made in 2008–2013, respectively.

Soil temperatures and water contents at a 5-cm depth were measured at a distance of 10 cm from the chamber collar at the same time as the SOC mineralization rates using a Li-Cor thermocouple probe and a Theta Probe ML2X with a HH2 water content meter (Delta-T Devices, Cambridge, England), respectively. Daily mean soil temperature and moisture data were provided by the State Key Agro-Ecological Experimental Station, both of which were measured at 5 cm below the surface using a Hydra soil moisture sensor (Hydra Data Reader and Hydra Probe II Soil Moisture Sensor (SDI-12/RS485); Precision: Moisture, ±0.5% vol; Temperature, ±0.6 ^{o}C; Stevens Water Monitoring Systems Inc., Australia). Soil water-filled pore space (WFPS) was calculated as follows: WFPS (%) = 100 × [volumetric water content / (2.65 − soil bulk density) / 2.65], with 2.65 being the particle density of the soil (g cm^{-3}).

2.4 Data analysis

An exponential (or "Q_{10}") function was used to simulate the relationship between SOC mineralization rate and soil temperature (Xu and Qi, 2001):

$$F = \beta_0 e^{\beta_1 T} \tag{1}$$

$$Q_{10} = e^{10\beta_1} \tag{2}$$

Where F (μ mol $m^{-2} s^{-1}$) is the SOC mineralization rate, T (℃) is the soil temperature at a depth of 5 cm, and β_0 and β_1 are the fitted parameters.

A quadratic polynomial function was used to simulate the relationship between SOC mineralization rate and soil moisture content (Tang et al., 2005):

$$F = \beta_3 \theta^2 + \beta_2 \theta + \beta_4 \tag{3}$$

Where θ is the soil moisture at a depth of 0–5 cm, and β_2, β_3, and β_4 are the fitted parameters.

The interactions of soil temperature with moisture content can more accurately simulate soil respiration than either soil temperature or moisture alone (Tang et al., 2005). Our data indicated that SOC mineralization rate increased with increasing soil moisture content to a maximum at approximately 46% WFPS, and then decreased with further increase of soil moisture content. After comparing different functions and resulting residual plots, a bivariate model was used to simulate the effect of soil moisture content and temperature on SOC mineralization rate:

$$F = \beta_0 e^{\beta_1 \mathrm{T}\theta + \beta_2 \mathrm{T}\theta^2} \tag{4}$$

The annual cumulative SOC mineralization rate was estimated by linear interpolating between measurement dates to obtain the mean daily SOC mineralization rate for each plot, and then summing the mean daily SOC mineralization rate for a given year.

The relationships between Q_{10} and meteorological factors were investigated using the SAS software (version 8.0; SAS Institute, Cary, NC). All other statistical analyses were performed with ANOVA at P = 0.05.

3. Results

3.1 Interannual variation in Q_{10}

The temporal variation in SOC mineralization rate was correlated with that of soil temperature in all six years (Figs. 2b and c), and it increased exponentially with soil temperature (P<0.01). The mean annual SOC mineralization rate ranged from 0.83 (2012) to 1.22 μ mol $m^{-2} s^{-1}$ (2008), with a mean of 0.99 μ mol $m^{-2} s^{-1}$ and a CV of 17%; the annual cumulative SOC mineralization ranged from 226 (2012) to 298 g C $m^{-2} y^{-1}$ (2009), with a mean of 253 g C $m^{-2} y^{-1}$ and a CV of 13% (Table 2), and the annual Q_{10} in our sites was 1.65 in 2008, 1.94 in 2009, 1.72 in 2010, 1.48 in 2011, 1.86 in 2012, and 1.55 in 2013, respectively, with a mean Q_{10} of 1.72 and a CV of 10% (Table 3).

3.2 Interannual variation in soil microclimate

Annual precipitation showed a significant annual variation (Fig.1 and Table 2; P <0.05). Rainfall ranged from 481 (2009 and 2012) to 644 mm (2011), with a 6-year mean of 540 ±64 mm and a CV of 12%. Annual rainfall days ranged from 71 (2013) to 105 days (2008), with a 6-year mean of 96 ±12 days and a CV of 13%. Interannual variation in air temperature was not significant (Fig.1 and Table 2; P >0.05). It ranged from 9.43 (2011 and 2012) to 11.08 ^{o}C (2013), with a 6-year mean of 10.1 ±0.6 ^{o}C and a CV of only 6%.

Soil temperature and soil moisture at a depth of 0–5 cm showed significant temporal variations over the six-year observation period (Fig. 2b). The seasonal mean soil moisture content was 49.2% WFPS in the wet season (July to September in each year) and 38.6% WFPS in the dry season (other months). The mean annual soil moisture content ranged from 38.6% WFPS (2013) to 50.7% WFPS (2011), with a mean of 43.8% WFPS and a CV of 11%. The seasonal mean soil temperature was 14.50 ℃ in the dry season and 20.39 ℃ in the wet season. The mean annual soil temperature ranged from 14.90 ℃ (2011) to 18.42 ℃ (2009), with a mean of 17.05 ℃ and a CV of only 7%.

3.3 Effect of soil moisture on the interannual variation of Q_{10}

Annual Q_{10} showed a negative quadratic correlation with annual mean soil moisture (Fig. 3b). Additionally, the seasonal SOC mineralization rate increased exponentially with soil temperature, and showed a negative quadratic correlation with soil moisture content (Table 3). The response surface of SOC mineralization rate to soil temperature and moisture including both seasonal and interannual scales clearly described how soil microclimate influenced SOC mineralization rate (Fig. 4).

4. Discussion

4.1 Soil moisture influenced the interannual variation in Q_{10}

The range of annual Q_{10} (1.48–1.94, with a CV of 10%) in our sites for the period 2008–2013 was within the limits reported for annual Q_{10} (1.20–4.89) at global scale (Boone et al., 1998; Zhou et al., 2007; Gaumont-Guay et al., 2008; Zhu and Cheng, 2011; Zimmermann et al., 2012). However, the mean annual Q_{10} in our sites (1.70) was lower than the global mean (2.47) (Boone et al., 1998; Zhou et al., 2007; Gaumont-Guay et al., 2008; Zhu and Cheng, 2011; Zimmermann et al., 2012), probably due to low SOC contents, small microbial communities, dry soil conditions in semi-arid regions (Conant et al., 2004; Gershenson et al., 2009; Cable et al., 2011), and different methods used for separating SOC mineralization rate (Boone et al., 1998; Zhu and Cheng, 2011; Zimmermann et al., 2012).

Annual Q_{10} was negatively linearly correlated with annual mean precipitation, but this correlation did not reach statistical significance (P>0.05); whereas it was significantly related to soil moisture content (Fig. 3). This was in agreement with previous studies (Suseela et al., 2012; Poll et al., 2013). However, Q_{10} was found to be negatively correlated with mean annual precipitation (P<0.01) in different forest ecosystems in China, which could be due to the relatively abundant rainfall in the forest ecosystems (700–1956 mm) (Peng et al., 2009). Soil moisture was the major limiting factor for the underground biological processes, especially in water-limited regions (Reth et al., 2005; Balogh et al., 2011; Wang et al., 2014). Although precipitation was the only source of water for soil moisture underneath long-term bare soil, there was no significant relationship between annual mean soil moisture and annual precipitation amount (P>0.05) (Fig. 5a), but rainfall frequency and distribution were closely related to annual mean soil moisture content (Fig. 5b). Similar results have also been found in other studies (Coronato and Bertiller, 1996; Qiu et al., 2001; Cho and Choi, 2014). The annual precipitation during the six-year observation period of 2008–2013 ranged from 481 (2009) to 644 mm (2011), with a CV of 12% (Table 2). The annual mean soil moisture content was high (51% WFPS) in 2011 due to relatively uniform distribution of precipitation, and low (38% WFPS) in 2010 and 2013 due to relatively uneven distribution of precipitation. For example, the rainfall amount on July 23, 2010 (118 mm) and July 22, 2013 (121 mm) was about 20% and 23% of that in 2010 (588 mm) and 2013 (523 mm), respectively. The annual mean soil moisture was moderate (43–47% WFPS) in 2008, 2009 and 2012 due to the normal distribution of precipitation. Similarly, the interannual soil moisture regulation in the forest ecosystems in the Loess Plateau was determined not only by rainfall amount but also by rainfall distribution (Li et al., 1998).

Annual Q_{10} showed a negative quadratic relationship with soil moisture content, as it increased with increasing soil moisture content to a maximum at approximately 42% WFPS, and then decreased with further increase of soil moisture content (Fig. 3b), which was in agreement with other studies (Bowden et al., 1998; Conant et al., 2004; Smith, 2005). This could be attributed to the following reasons: Firstly, lower soil water availability could reduce Q_{10} by limiting respiration substrate availability and soil pore water became increasingly disconnected, thus slowing down the diffusion rate of solutes (Wan et al., 2007; Balogh et al., 2011), and decreasing the activity and quantity of organisms due to drought stress (Davidson et al., 2006). Secondly, higher soil moisture could also reduce Q_{10} by limiting O_2 diffusion rate (Davidson et al., 1998; Byrne et al., 2005; Saiz et al., 2007) because of low effective soil porosity, as the diffusion rate of O_2 through water was much slower than that through air (Cook and Knight, 2003; Manzoni et al., 2012), thus the decomposition activity of aerobic microbes was inhibited due to lack of oxygen (Davidson et al., 2000). Finally, the diffusion rate of both soluble organic matter and O_2 were not inhibited, also the survival of microorganisms not subject to water stress at suitable soil water content, instead increasing temperature increased the diffusion of soluble organic matter, thus resulting in an increase in Q_{10} (McCulley et al., 2007). Overall, soil moisture content may be the most important factors that affected the interannual variation in Q_{10}.

The variation in the temperature sensitivities of SOC mineralization could have potential implications for climate carbon modeling (Davidson and Janssens, 2006; Conant et al., 2011), as uncertainty remains regarding environmental controls over SOC mineralization (Larionova et al., 2007; Karhu et al., 2010; Conant et al., 2011; Sakurai et al., 2012). The previous results have emphasized the importance of seasonal variation in precipitation and soil moisture in determining Q_{10} (Xu and Qi, 2001; Davidson et al., 2006; Davidson and Janssens, 2006), but have rarely taken into account the interannual variation in soil moisture resulting from the uneven distribution of precipitation. Carbon cycle modeling without considering this interannual variation in soil moisture may produce misleading conclusions.

4.2 Comparison with annual cumulative SOC mineralization rate estimated by different methods

Annual cumulative SOC mineralization rate was estimated by different methods, including linear interpolation method, modeled method, and unit conversion method. The results clearly showed that there was no significant difference in the estimates of annual cumulative SOC mineralization rate between linear interpolation and modeled method, and the modeled method could well predict the SOC mineralization rate in most cases from 2008 to 2013 (Fig. 6), which was in line with the previous studies (Tang et al., 2005). However, unit conversion method seriously overestimated annual cumulative SOC mineralization rate (Table 4). This can be attributed to the following reasons: 1) the study site has a continental monsoon climate with 60% of rainfall occurring from July to September (rainy season), thus the study site is hot and rainy in the rainy season, but cool and dry in the non-rainy season; and 2) SOC mineralization rate in the rainy and non-rainy season is largely the same, but the duration of rainy season is only a quarter of a year. Thus, the SOC mineralization rate was much greater in rainy season than in non-rainy season, thus resulting in an overestimation of cumulative SOC mineralization rate in a given year.

In conclusion, linear interpolation method is a simple and controllable method for estimating annual cumulative SOC mineralization rate (Schindlbacher et al., 2014; Shi et al., 2014). Although the modeled method can well estimate annual cumulative SOC mineralization rate, it is limited in practice as it needs daily soil temperature and moisture. Unit conversion method may seriously overestimate annual cumulative SOC mineralization rate unless the SOC mineralization rate is very uniform in a given year.

5. Conclusions

Understanding the factors influencing the temperature sensitivity of SOC mineralization is important to accurately estimate local carbon cycle. The results of this study showed that the annual cumulative SOC mineralization ranged from 226 to 298 g C m^{-2} y^{-1}, with a CV of 13%, annual Q_{10} ranged from 1.48 to 1.94, with a CV of 10%, and annual soil moisture content ranged from 38.6 to 50.7% WFPS, with a CV of 11%. Annual Q_{10} showed a negative quadratic correlation with annual mean soil moisture, which was determined by uneven distribution and frequency of rainfall. In conclusion, the interannual variation in soil moisture content should be considered in carbon cycle models in semi-arid areas.

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Table 1. SOC mineralization rate (μ mol m^{-2} s^{-1}) in summer (July 11, 2008) and winter (November 18, 2008). Data are represented as mean ± S.D of five collars.

	SOC mineralization rate					
Dates	Collar 1	Collar 2	Collar 3	Collar 4	Collar 5	Mean value
Summer	1.55 ±0.11	1.60 ±0.20	1.58 ±0.21	1.49 ±0.07	1.65 ±0.18	1.57 ±0.06
Winter	0.29 ±0.01	0.30 ±0.02	0.31 ±0.01	0.32 ±0.02	0.33 ±0.02	0.31 ±0.02

Note: SOC mineralization rate was measured on July 11, 2008 and November 18, 2008 (representing summer and winter) using 5 PVC collars installed in our plots

Table 2. Cumulative SOC mineralization rate (g C m^{-2} $year^{-1}$), annual precipitation amount (mm), annual precipitation days, and air temperature (^{o}C) from 2009 to 2013. Data are represented as mean ± S.D.

Years	Cumulative SOC mineralization rate	Precipitation amount	Precipitation days	Air temperature
2008	293±10	520	105	9.76
2009	298±9	481	99	10.26
2010	238±50	588	101	10.39
2011	234±48	644	100	9.43
2012	226±19	481	98	9.43
2013	240±30	523	71	11.08
Mean	253±32	540±64	96±12	10.1±0.6

Table 3. Relationships between SOC mineralization rate and soil temperature (F-T) or soil moisture (F-θ) for each year from 2008 to 2013.

Years	F-T				F-θ		
	Functions	R^2	P	Q_{10}	Functions	R^2	P
2008	$F=0.49e^{0.0499T}$	0.56	<0.01	1.65	$F=-0.0008\theta^2 + 0.10\theta - 1.52$	0.53	<0.01
2009	$F=0.34e^{0.0661T}$	0.63	<0.01	1.94	$F=-0.0001\theta^2 - 0.02\theta + 2.63$	0.61	<0.01
2010	$F=0.35e^{0.0544T}$	0.47	<0.01	1.72	$F=0.0002\theta^2 - 0.04\theta + 2.15$	0.86	<0.01
2011	$F=0.45e^{0.0395T}$	0.47	<0.01	1.48	$F=-0.0008\theta^2 + 0.06\theta + 0.06$	0.46	<0.01
2012	$F=0.27e^{0.0623T}$	0.67	<0.01	1.86	$F=-0.0019\theta^2 + 0.14\theta - 1.71$	0.35	<0.05
2013	$F=0.52e^{0.0441T}$	0.32	<0.01	1.55	$F=-0.001\theta^2 + 0.08\theta - 0.60$	0.36	<0.05

Table 4. Annual cumulative SOC mineralization rate (g C m^{-2} $year^{-1}$) estimated by linear interpolation method, modeled method, and unit conversed method from 2008 to 2013.

Years	Annual cumulative SOC mineralization rate		
	Linear interpolation	Soil temperature and moisture modeled	Unit conversion
2008	293	258	462
2009	298	272	460
2010	238	268	344
2011	234	260	325
2012	226	271	314
2013	240	284	348
Mean	255±32	269±6	374±65

Note: Modeled method: using the interactions of soil temperature with moisture for estimating annual cumulative SOC mineralization rate with Eq. 4 (2.4 sections); Unit conversion method: estimating annual cumulative SOC mineralization rate with mean SOC mineralization rate in a given year.

Figure captions

Fig. 1

Location of the State Key Agro-Ecological Experimental Station (Changwu Station).

Fig. 2

Temporal variations of (a) precipitation and air temperature, (b) soil moisture and soil temperature, and (c) SOC mineralization rate from 2008 to 2013.

Fig. 3

Regression analysis performed between (a) Q_{10} and annual precipitation amount, and (b) Q_{10} and annual mean soil moisture.

Fig. 4

Response surface of SOC mineralization rate as a function of soil moisture and soil temperature from 2008 to 2013.

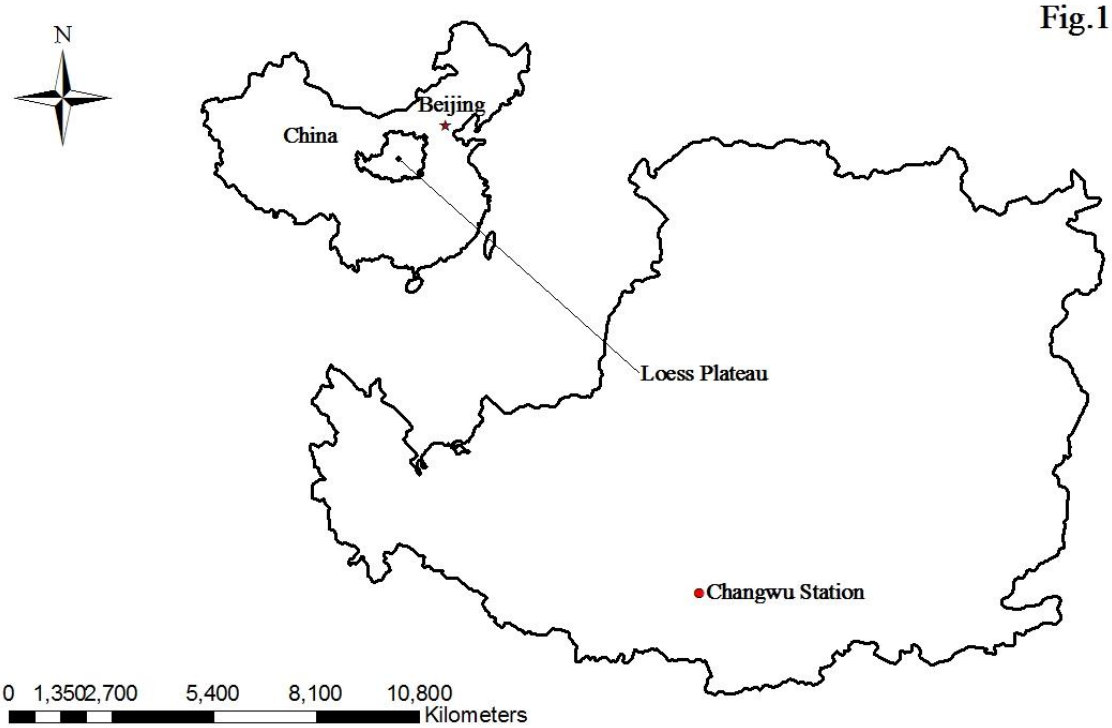
Fig. 5

Regression analysis performed between (a) annual mean soil moisture and annual precipitation amount, and (b) annual mean soil moisture and annual precipitation days.

Fig. 6

Estimated daily (2008–2013) SOC mineralization rate (solid line) with periodic measurement values (filled circles).

Fig.1


N
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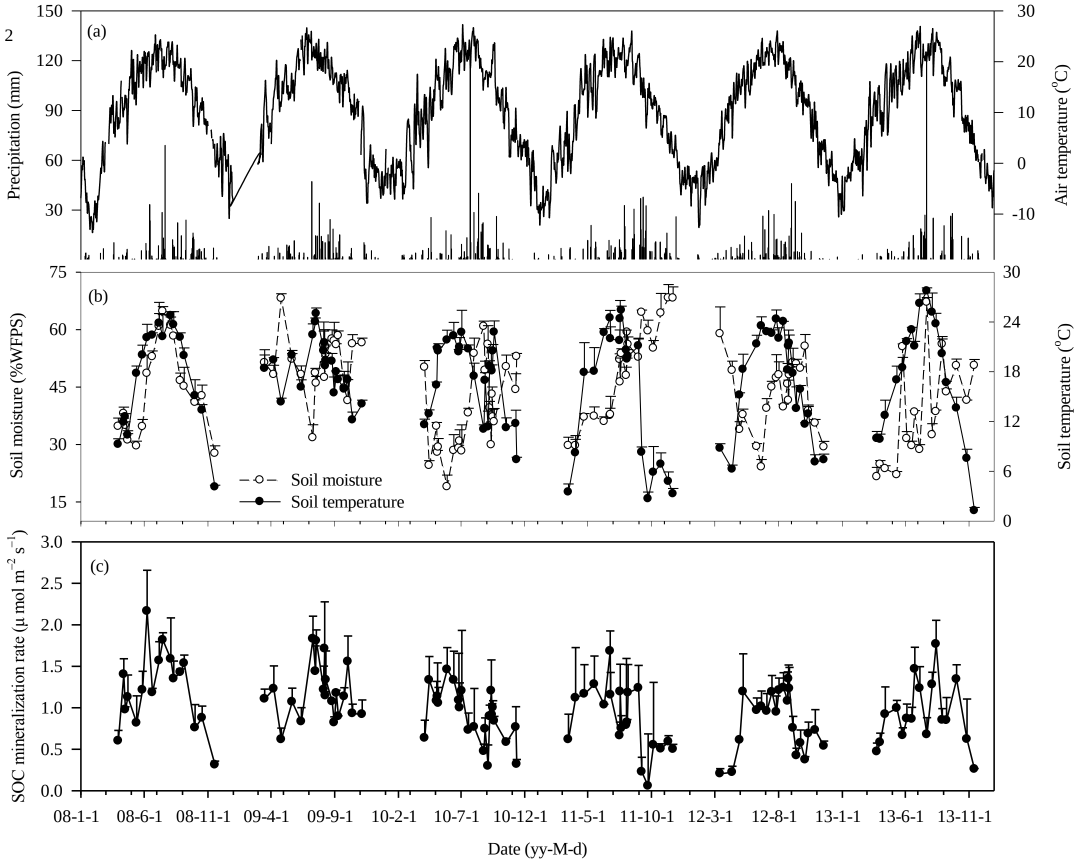
China
Loess Plateau
Changwu Station
0 1,350 2,700 5,400 8,100 10,800
Kilometers

Fig. 2



(a)
Precipitation (mm)
Air temperature (°C)
(b)
Soil moisture (%WFPS)
Soil temperature (°C)
Soil moisture
Soil temperature
(c)
SOC mineralization rate (μ mol m⁻² s⁻¹)
08-1-1
08-6-1
08-11-1
09-4-1
09-9-1
10-2-1
10-7-1
10-12-1
11-5-1
11-10-1
12-3-1
12-8-1
13-1-1
13-6-1
13-11-1
Date (yy-M-d)

601

Fig. 3

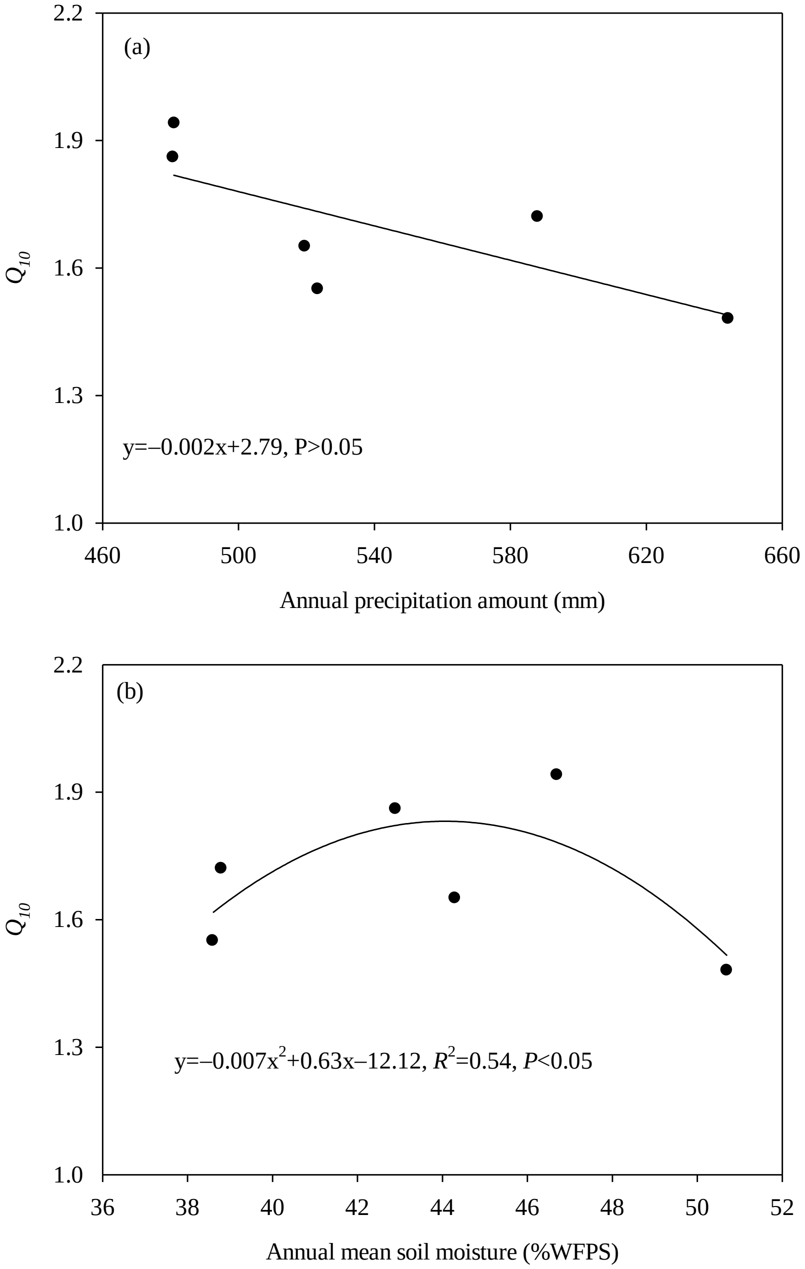
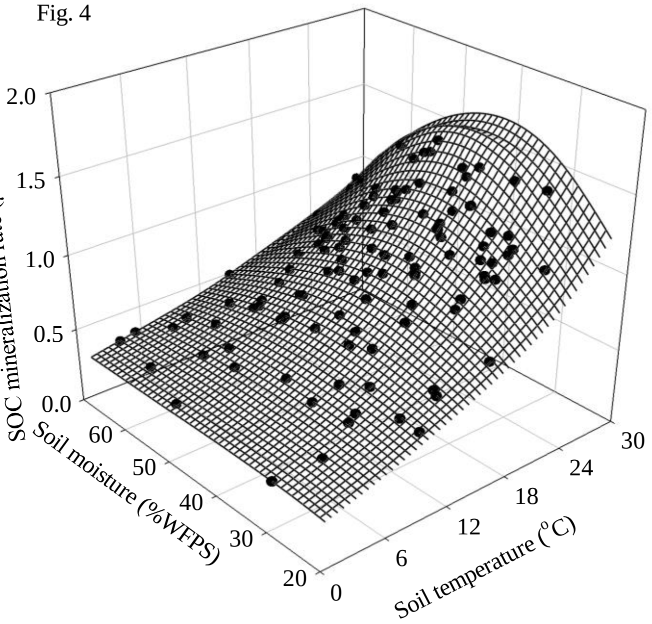

(a)
2.2
1.9
1.6
1.3
1.0
Q10
y=–0.002x+2.79, P>0.05
460
500
540
580
620
660
Annual precipitation amount (mm)
(b)
2.2
1.9
1.6
1.3
1.0
Q10
y=–0.007x2+0.63x–12.12, R2=0.54, P<0.05
36
38
40
42
44
46
48
50
52
Annual mean soil moisture (%WFPS)

Fig. 4


SOC mineralization rate (μ mol m^{-2} s^{-1})
2.0
1.5
1.0
0.5
0.0
Soil moisture (%WFPS)
60
50
40
30
20
Soil temperature (^{o}C)
0
6
12
18
24
30

604

Fig.5

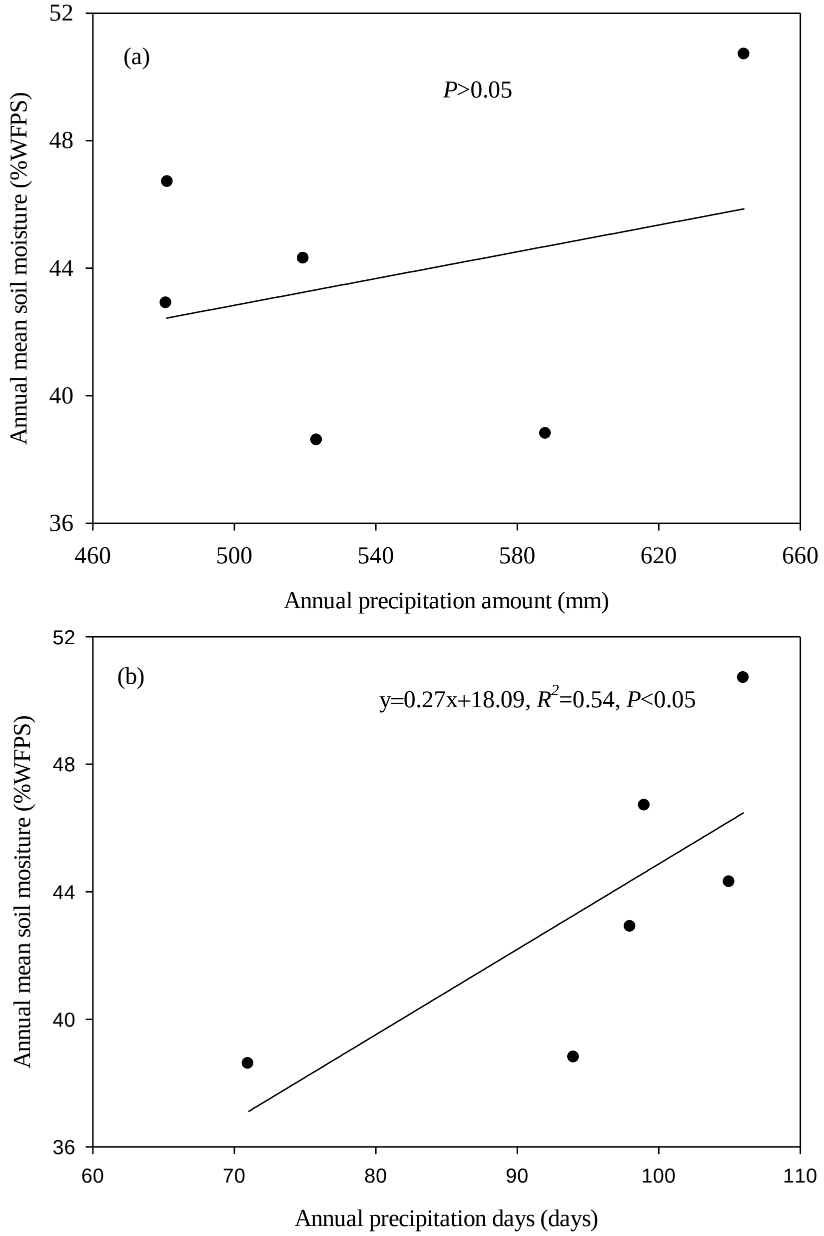
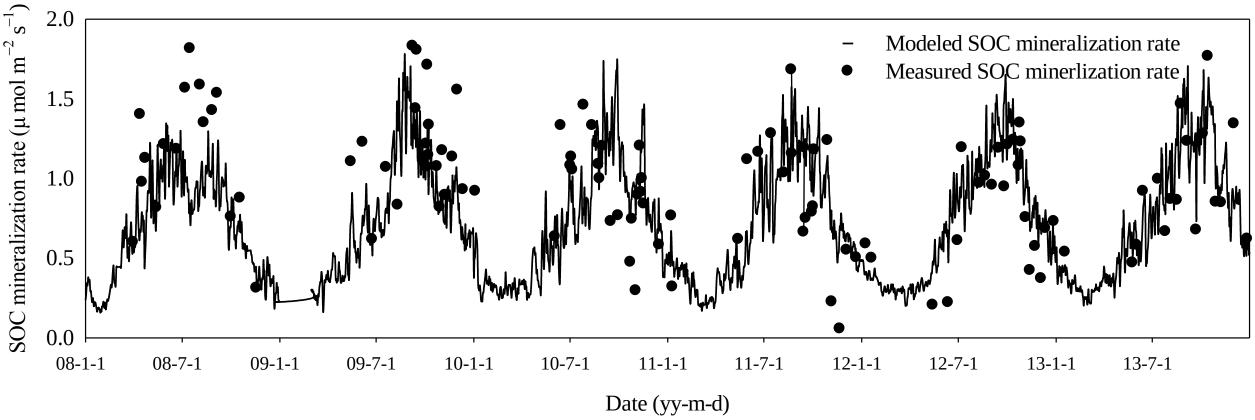

(a)
P>0.05
Annual mean soil moisture (%WFPS)
52
48
44
40
36
460
500
540
580
620
660
Annual precipitation amount (mm)
(b)
y=0.27x+18.09, R2=0.54, P<0.05
Annual mean soil mositure (%WFPS)
52
48
44
40
36
60
70
80
90
100
110
Annual precipitation days (days)

Fig. 6



SOC mineralization rate (μ mol m^{-2} s^{-1})
2.0
1.5
1.0
0.5
0.0
Modeled SOC mineralization rate
Measured SOC minerlization rate
08-1-1
08-7-1
09-1-1
09-7-1
10-1-1
10-7-1
11-1-1
11-7-1
12-1-1
12-7-1
13-1-1
13-7-1
Date (yy-m-d)

617