

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

4

5 Y.J Zhang<sup>1,2</sup>, S.L Guo<sup>1,3,4\*</sup>, M Zhao<sup>4</sup>, L.L Du<sup>1</sup>, R.J Li<sup>1</sup>, J.S Jiang<sup>3</sup>, R Wang<sup>4</sup>, and N.N Li<sup>1</sup>

6 **Affiliation:**

7 1. State Key Laboratory of Soil Erosion and Dry-land Farming on the Loess Plateau, Institute  
8 of Soil and Water Conservation, Northwest A&F University, Yangling 712100, China

9 2. Geography and Environmental Engineering Department, Baoji University of Arts and  
10 Sciences, Baoji 721013, China

11 3. Institute of Soil and Water Conservation, Chinese Academy of Sciences and Ministry of  
12 Water Resource, Yangling 712100, China

13 4. College of Resources and Environment, Northwest A&F University, Yangling 712100,  
14 China

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19 Corresponding author: Shengli Guo

20 Address: Institute of Soil and Water Conservation, Xinong Road 26, Yangling, Shaanxi  
21 712100, China

22 Phone: +86-29-87012411, Fax: +86-29-87012210, E-mail: [slguo@ms.iswc.ac.cn](mailto:slguo@ms.iswc.ac.cn)

23 **Abstract**

24 Temperature sensitivity of soil organic carbon (SOC) mineralization (i.e.,  $Q_{10}$ )  
25 determines how strong the feedback from global warming may be on the atmospheric CO<sub>2</sub>  
26 concentration, thus understanding the factors influencing the interannual variation in  $Q_{10}$  is  
27 important to accurately estimate local soil carbon cycle. *In situ* SOC mineralization rate was  
28 measured using an automated CO<sub>2</sub> flux system (Li-8100) in long-term bare fallow soil in the  
29 Loess Plateau (35°12' N, 107°40' E) in Changwu, Shaanxi, China from 2008 to 2013. The  
30 results showed that the annual cumulative SOC mineralization ranged from 226 to 298 g C  
31 m<sup>-2</sup> y<sup>-1</sup>, with a mean of 253 g C m<sup>-2</sup> y<sup>-1</sup> and a CV of 13%, annual  $Q_{10}$  ranged from 1.48 to  
32 1.94, with a mean of 1.70 and a CV of 10%, and annual soil moisture content ranged from  
33 38.6 to 50.7% soil water-filled pore space (WFPS), with a mean of 43.8% WFPS and a CV of  
34 11%, which were mainly affected by the frequency and distribution of precipitation. Annual  
35  $Q_{10}$  showed a quadratic correlation with annual mean soil moisture content. In conclusion,  
36 understanding of the relationships between interannual variation in  $Q_{10}$ , soil moisture and  
37 precipitation are important to accurately estimate the local carbon cycle, especially under the  
38 changing climate.

39

40

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

43

44 **1. Introduction**

45 Temperature sensitivity of soil organic carbon (SOC) mineralization (hereafter refer to  
46 as  $Q_{10}$ ) is of critical importance because it determines how strong the feedback from global  
47 warming may be on the atmospheric CO<sub>2</sub> concentration (Ågren and Wetterstedt, 2007).  
48 However, this is an issue of considerable debatable (Davidson et al., 2006; Kirschbaum,  
49 2006), and the variations in  $Q_{10}$  are the main source of controversies in this feedback intensity  
50 (Larionova et al., 2007; Karhu et al., 2010; Conant et al., 2011; Sakurai et al., 2012).  
51 Therefore, understanding the factors influencing  $Q_{10}$  is important to accurately estimate C  
52 cycle and the feedback from the expected warmer climate.

53 Previous studies have shown that  $Q_{10}$  variations are closely related to soil temperature  
54 (Kirschbaum, 2006; Von Lutzow and Kogel-Knabner, 2009), substrate availability (Ågren  
55 and Wetterstedt, 2007; Gershenson et al., 2009), substrate quality (Von Lutzow and  
56 Kogel-Knabner, 2009; Sakurai et al., 2012), and the size and composition of microbial  
57 population (Djukic et al., 2010; Karhu et al., 2010). Soil moisture is the most significant  
58 limiting factor for underground physiological processes in dry and semi-dry ecosystems  
59 (Balogh et al., 2011; Cable et al., 2011; Wang et al., 2014). Soil water availability may  
60 indirectly affect  $Q_{10}$  by influencing the diffusion of substrates, because the diffusion of  
61 extracellular enzymes produced by microorganisms and available substrates must conduct in  
62 the liquid phase (Davidson et al., 1998; Illeris et al., 2004), but the response of  $Q_{10}$  to soil  
63 water availability is extremely complex and controversial (Davidson et al., 2000; Davidson et  
64 al., 2006; McCulley et al., 2007). For example, Guldge and Schimel (2000) found that  $Q_{10}$   
65 was larger in wet years than in drought years, whereas the opposite result was found by Dorr  
66 and Mdnich (1987). However, many other studies that mainly focused on the short-term or  
67 seasonal variation in  $Q_{10}$  (Davidson et al., 2006) have showed that  $Q_{10}$  was not affected by  
68 soil moisture (Fang and Moncrieff, 2001; Reichstein et al., 2002; Jassal et al., 2008).

69 Additionally, soil water availability experienced marked seasonal and interannual fluctuations  
70 in these ecosystems due to uneven rainfall distribution caused by the abnormal increase of  
71 atmospheric CO<sub>2</sub> concentrations (Solomon et al., 2007). The uneven rainfall distribution  
72 inevitably influenced soil moisture availability (Coronato and Bertiller, 1996; Qiu et al., 2001;  
73 Cho and Choi, 2014). Xiao et al. (2014) have shown that the interannual changes in soil  
74 moisture storage in the Loess Plateau were decided by the difference in soil moisture storage  
75 between October and April, because precipitation from April to October of 2004 to 2010  
76 accounted for at least 86% of annual rainfall. However, to our knowledge, there have been  
77 few studies investigating the relationship between interannual variation in  $Q_{10}$  and soil  
78 moisture under natural conditions.

79 The Loess Plateau is located in northwest China covering an area of 640,000 km<sup>2</sup>. It has  
80 a continental monsoonal climate and shows a dramatically interannual fluctuations in  
81 precipitation, with the highest precipitation of 1262 mm and the lowest precipitation of only  
82 80 mm, and a mean value of 150–750 mm (Lin and Wang, 2007). The precipitation in the  
83 loess regions also shows a dramatically seasonal variation, and approximately 60%–80% of  
84 the annual precipitation falls during the three summer months from July to September (Guo et  
85 al., 2012). Several recent studies have attempted to determine the dominant factors  
86 responsible for the variation of soil respiration in vegetation ecosystems (Lafond et al., 2011;  
87 Shi et al., 2011; Jurasic et al., 2012). However, there have been no studies on the  
88 interannual variation in  $Q_{10}$ , nor the factors responsible for these changes. This highlights the  
89 need to accurately evaluate the response of SOC mineralization to increasing temperature  
90 under warmer climate scenarios in the eroded or degraded regions, because air temperature  
91 has been increasing over the past decades (Fan and Wang, 2011; Wang et al., 2012). Thus, the  
92 objectives of the present study are to (1) quantify the interannual variation in  $Q_{10}$ ; (2)  
93 determine the effect of soil moisture on this interannual variation for the period 2008–2013 in

94 the Loess Plateau, China.

95

96 **2. Materials and methods**

97 **2.1 Site description**

98 This study was a part of a long-term field experiment that began in 1984 in the State  
99 Key Agro-Ecological Experimental Station in the Loess Plateau in Changwu, Shaanxi, China  
100 (35°12' N, 107°40' E; 1,200 m above sea level) (Fig. 1). This region had a continental  
101 monsoon climate with a mean annual precipitation of 560 mm for the period 1984–2013,  
102 over 60% of which occurred from July to September. During this 30-year period, the annual  
103 mean air temperature was 9.4 °C and the monthly mean temperature between July and  
104 September was 19.4 °C. The study site is also characterized by a  $\geq 10$  °C accumulated  
105 temperature of 3029 °C, an annual sunshine duration of 2230 h, an annual total radiation of  
106 484 kJ cm<sup>-2</sup>, and a frost-free period of 171 days.

107 The site was located in a typical rain-fed cropping region of the Loess Plateau highland  
108 in northwest China. The soil was classified as a loam (Cumulic Haplustoll, USDA Soil  
109 Taxonomy System) developed from loess deposits. Soils collected at the study site in 1984 at  
110 a depth of 0–20 cm contained 10.5% CaCO<sub>3</sub>, 6.5 g organic C kg<sup>-1</sup>, 0.80 g total N kg<sup>-1</sup>, and  
111 200 mg NH<sub>4</sub>OAc-extractable K kg<sup>-1</sup>, 3.0 g kg<sup>-1</sup> available phosphorus, and had a pH of 8.4  
112 (with a 1: 1 ratio of soil: H<sub>2</sub>O), a water-holding capacity of 0.29 cm<sup>3</sup> cm<sup>-3</sup> (v/v), the wilting  
113 point of 11%, a soil bulk density of 1.3 g cm<sup>-3</sup>, soil porosity of 51%, and a clay content of  
114 24%.

115 **2.2 Experimental design and management**

116 This study was a part of a long-term field experiment established in June 1984. The plot  
117 used in the present study is taken from a bare plot in a state of fallow since June 1984 after  
118 the harvesting of winter wheat (*Triticum aestivum* L. 'Chang Wu 131 series'), and living

119 weed was artificially removed timely. Therefore, there were no vegetation or inputs of  
120 aboveground and belowground litter, and then SOC mineralization rates in the bare fallow  
121 soil did not include root respiration and litter mineralization and decomposition. In this paper,  
122 three bare fallow plots were used to investigate the mechanism of underground SOC  
123 mineralization rates. All plots of  $10.3\text{ m} \times 6.5\text{ m}$  ( $66.95\text{ m}^2$ ) were randomly arranged in three  
124 blocks. The plots were separated by  $0.5\text{ m}$  spaces, whereas the blocks were separated by  $1\text{ m}$   
125 strips.

126

### 127 **2.3 Measurements of SOC mineralization rate and soil microclimate**

128 SOC mineralization rate was measured using an automated closed soil  $\text{CO}_2$  flux system  
129 with a portable chamber (20 cm in diameter, Li-8100, Lincoln, NE, USA). Approximately  
130 one day before the first measurement, a polyvinyl chloride (PVC) collar (20 cm in diameter  
131 and 12 cm in height) was inserted to a depth of 2 cm into each plot, and left in place  
132 throughout the experimental period from 2008 to 2013. Five PVC collars were installed in  
133 our plots for investigating the spatial variation of SOC mineralization rate in summer (11 July  
134 2008) and winter (18 November 2008), respectively. Although previous studies have  
135 demonstrated a significant spatial variation of soil respiration, especially in the sites with  
136 complex terrain (causing the redistribution of SOC) and different vegetation types (Epron et  
137 al., 2006; Luan et al., 2012), the spatial variation of SOC mineralization rate in our sites is  
138 small with a variation coefficient of only 4% and 5% in summer and winter, respectively  
139 (Table 1). The results implied that the temporal fluctuation during the measurement have little  
140 effect on the spatial variation of SOC mineralization rate. This can be attributed to that there  
141 have been no vegetation or inputs of (aboveground and belowground) litter in our plots since  
142 1984 (absolute fallow), and the soil was derived aeolian deposit loess and flat terrain. Due to  
143 the small areas of our plots ( $66.95\text{ m}^2$ ) and time constraints (5 min for measuring SOC

144 mineralization rate in a given PVC collar), only one PVC collar was used in each plot for  
145 measuring SOC mineralization rate and only twice measures were conducted for studying the  
146 spatial variation of SOC mineralization rate in our plots. All visible living organisms were  
147 removed before the measurement. If necessary, one or more additional measurements would  
148 be taken until the variations between two consecutive measurements were less than 15%. The  
149 final instantaneous soil respiration for a given collar was the average of the two  
150 measurements with a 90 s enclosure period and 30 s delay between them. Field measurements  
151 were performed between 09:00 and 11:00 AM from March 2008 to November 2013, except  
152 in December, January, and February because of cold weather. A total of 17, 25, 26, 22, 26 and  
153 17 SOC mineralization measurements were made in 2008–2013, respectively.

154 Soil temperatures and water contents at a 5-cm depth were measured at a distance of 10  
155 cm from the chamber collar at the same time as the SOC mineralization rates using a Li-Cor  
156 thermocouple probe and a Theta Probe ML2X with a HH2 water content meter (Delta-T  
157 Devices, Cambridge, England), respectively. Daily mean soil temperature and moisture data  
158 were provided by the State Key Agro-Ecological Experimental Station, both of which were  
159 measured at 5 cm below the surface using a Hydra soil moisture sensor (Hydra Data Reader  
160 and Hydra Probe II Soil Moisture Sensor (SDI-12/RS485); Precision: Moisture,  $\pm 0.5\%$  vol;  
161 Temperature,  $\pm 0.6\text{ }^{\circ}\text{C}$ ; Stevens Water Monitoring Systems Inc., Australia). Soil water-filled  
162 pore space (WFPS) was calculated as follows:  $\text{WFPS} (\%) = 100 \times [\text{volumetric water content} /$   
163  $(2.65 - \text{soil bulk density}) / 2.65]$ , with 2.65 being the particle density of the soil ( $\text{g cm}^{-3}$ ).

164

## 165 **2.4 Data analysis**

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

$$168 F = \beta_0 e^{\beta_1 T} \quad (1)$$

$$Q_{l0} = e^{l0\beta_l} \quad (2)$$

170 Where  $F$  ( $\mu \text{ mol m}^{-2} \text{ s}^{-1}$ ) is the SOC mineralization rate,  $T$  ( $^{\circ}\text{C}$ ) is the soil temperature at a  
 171 depth of 5 cm, and  $\beta_0$  and  $\beta_1$  are the fitted parameters.

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

$$F = \beta_3 \theta^2 + \beta_2 \theta + \beta_1 \quad (3)$$

175 Where  $\theta$  is the soil moisture at a depth of 0–5 cm, and  $\beta_2, \beta_3$ , and  $\beta_4$  are the fitted parameters.

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

$$F = \beta_0 e^{\beta_1 T \theta + \beta_2 T \theta^2} \quad (4)$$

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

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

190

191 3. Results

### 192 3.1 Interannual variation in $O_{10}$

193 The temporal variation in SOC mineralization rate was correlated with that of soil

194 temperature in all six years (Figs. 2b and c), and it increased exponentially with soil  
195 temperature ( $P<0.01$ ). The mean annual SOC mineralization rate ranged from 0.83 (2012) to  
196  $1.22 \mu \text{ mol m}^{-2} \text{ s}^{-1}$  (2008), with a mean of  $0.99 \mu \text{ mol m}^{-2} \text{ s}^{-1}$  and a CV of 17%; the annual  
197 cumulative SOC mineralization ranged from 226 (2012) to 298  $\text{g C m}^{-2} \text{ y}^{-1}$  (2009), with a  
198 mean of  $253 \text{ g C m}^{-2} \text{ y}^{-1}$  and a CV of 13% (Table 2), and the annual  $Q_{10}$  in our sites was 1.65  
199 in 2008, 1.94 in 2009, 1.72 in 2010, 1.48 in 2011, 1.86 in 2012, and 1.55 in 2013,  
200 respectively, with a mean  $Q_{10}$  of 1.72 and a CV of 10% (Table 3).

201

### 202 **3.2 Interannual variation in soil microclimate**

203 Annual precipitation showed a significant annual variation (Fig.1 and Table 2;  $P <0.05$ ).  
204 Rainfall ranged from 481 (2009 and 2012) to 644 mm (2011), with a 6-year mean of  $540 \pm 64$   
205 mm and a CV of 12%. Annual rainfall days ranged from 71 (2013) to 105 days (2008), with a  
206 6-year mean of  $96 \pm 12$  days and a CV of 13%. Interannual variation in air temperature was  
207 not significant (Fig.1 and Table 2;  $P >0.05$ ). It ranged from 9.43 (2011 and 2012) to 11.08 °C  
208 (2013), with a 6-year mean of  $10.1 \pm 0.6$  °C and a CV of only 6%.

209 Soil temperature and soil moisture at a depth of 0–5 cm showed significant temporal  
210 variations over the six-year observation period (Fig. 2b). The seasonal mean soil moisture  
211 content was 49.2% WFPS in the wet season (July to September in each year) and 38.6%  
212 WFPS in the dry season (other months). The mean annual soil moisture content ranged from  
213 38.6% WFPS (2013) to 50.7% WFPS (2011), with a mean of 43.8% WFPS and a CV of 11%.  
214 The seasonal mean soil temperature was 14.50 °C in the dry season and 20.39 °C in the wet  
215 season. The mean annual soil temperature ranged from 14.90 °C (2011) to 18.42 °C (2009),  
216 with a mean of 17.05 °C and a CV of only 7%.

217

### 218 **3.3 Effect of soil moisture on the interannual variation of $Q_{10}$**

219 Annual  $Q_{10}$  showed a negative quadratic correlation with annual mean soil moisture (Fig.  
220 3b). Additionally, the seasonal SOC mineralization rate increased exponentially with soil  
221 temperature, and showed a negative quadratic correlation with soil moisture content (Table 3).  
222 The response surface of SOC mineralization rate to soil temperature and moisture including  
223 both seasonal and interannual scales clearly described how soil microclimate influenced SOC  
224 mineralization rate (Fig. 4).

225

## 226 **4. Discussion**

### 227 **4.1 Soil moisture influenced the interannual variation in $Q_{10}$**

228 The range of annual  $Q_{10}$  (1.48–1.94, with a CV of 10%) in our sites for the period  
229 2008–2013 was within the limits reported for annual  $Q_{10}$  (1.20–4.89) at global scale (Boone  
230 et al., 1998; Zhou et al., 2007; Gaumont-Guay et al., 2008; Zhu and Cheng, 2011;  
231 Zimmermann et al., 2012). However, the mean annual  $Q_{10}$  in our sites (1.70) was lower than  
232 the global mean (2.47) (Boone et al., 1998; Zhou et al., 2007; Gaumont-Guay et al., 2008;  
233 Zhu and Cheng, 2011; Zimmermann et al., 2012), probably due to low SOC contents, small  
234 microbial communities, dry soil conditions in semi-arid regions (Conant et al., 2004;  
235 Gershenson et al., 2009; Cable et al., 2011), and different methods used for separating SOC  
236 mineralization rate (Boone et al., 1998; Zhu and Cheng, 2011; Zimmermann et al., 2012).

237 Annual  $Q_{10}$  was negatively linearly correlated with annual mean precipitation, but this  
238 correlation did not reach statistical significance ( $P>0.05$ ); whereas it was significantly related  
239 to soil moisture content (Fig. 3). This was in agreement with previous studies (Suseela et al.,  
240 2012; Poll et al., 2013). However,  $Q_{10}$  was found to be negatively correlated with mean  
241 annual precipitation ( $P<0.01$ ) in different forest ecosystems in China, which could be due to  
242 the relatively abundant rainfall in the forest ecosystems (700–1956 mm) (Peng et al., 2009).  
243 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 23 July 2010 (118 mm) and 22 July 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<sub>2</sub> diffusion

269 rate (Davidson et al., 1998; Byrne et al., 2005; Saiz et al., 2007) because of low effective soil  
270 porosity, as the diffusion rate of O<sub>2</sub> through water was much slower than that through air  
271 (Cook and Knight, 2003; Manzoni et al., 2012), thus the decomposition activity of aerobic  
272 microbes was inhibited due to lack of oxygen (Davidson et al., 2000). Finally, the diffusion  
273 rate of both soluble organic matter and O<sub>2</sub> were not inhibited, also the survival of  
274 microorganisms not subject to water stress at suitable soil water content, instead increasing  
275 temperature increased the diffusion of soluble organic matter, thus resulting in an increase in  
276  $Q_{10}$  (McCulley et al., 2007). Overall, soil moisture content may be the most important factors  
277 that affected the interannual variation in  $Q_{10}$ .

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

287

288 **4.2 Comparison with annual cumulative SOC mineralization rate estimated by different  
289 methods**

290 Annual cumulative SOC mineralization rate was estimated by different methods,  
291 including linear interpolation method, modeled method, and unit conversion method. The  
292 results clearly showed that there was no significant difference in the estimates of annual  
293 cumulative SOC mineralization rate between linear interpolation and modeled method, and

294 the modeled method could well predict the SOC mineralization rate in most cases from 2008  
295 to 2013 (Fig. 6), which was in line with the previous studies (Tang et al., 2005). However,  
296 unit conversion method seriously overestimated annual cumulative SOC mineralization rate  
297 (Table 4). This can be attributed to the following reasons: 1) the study site has a continental  
298 monsoon climate with 60% of rainfall occurring from July to September (rainy season), thus  
299 the study site is hot and rainy in the rainy season, but cool and dry in the non-rainy season;  
300 and 2) SOC mineralization rate in the rainy and non-rainy season is largely the same, but the  
301 duration of rainy season is only a quarter of a year. Thus, the SOC mineralization rate was  
302 much greater in rainy season than in non-rainy season, thus resulting in an overestimation of  
303 cumulative SOC mineralization rate in a given year.

304 In conclusion, linear interpolation method is a simple and controllable method for  
305 estimating annual cumulative SOC mineralization rate (Schindlbacher et al., 2014; Shi et al.,  
306 2014). Although the modeled method can well estimate annual cumulative SOC  
307 mineralization rate, it is limited in practice as it needs daily soil temperature and moisture.  
308 Unit conversion method may seriously overestimate annual cumulative SOC mineralization  
309 rate unless the SOC mineralization rate is very uniform in a given year.

310

## 311 **5. Conclusions**

312 Understanding the factors influencing the temperature sensitivity of SOC mineralization  
313 is important to accurately estimate local carbon cycle. The results of this study showed that  
314 the annual cumulative SOC mineralization ranged from 226 to 298 g C m<sup>-2</sup> y<sup>-1</sup>, with a CV of  
315 13%, annual  $Q_{10}$  ranged from 1.48 to 1.94, with a CV of 10%, and annual soil moisture  
316 content ranged from 38.6 to 50.7% WFPS, with a CV of 11%. Annual  $Q_{10}$  showed a negative  
317 quadratic correlation with annual mean soil moisture, which was determined by uneven  
318 distribution and frequency of rainfall. In conclusion, the interannual variation in soil moisture

319 content should be considered in carbon cycle models in semi-arid areas.

320

321 **Author contribution.**

322 S.L Guo and M Zhao conceived and designed the experiments, R Wang and N.N Li  
323 performed the experiments, L.L Du and J.S Jiang analyzed the data, Y.J Zhang and R.J Li  
324 wrote the paper.

325

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330

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480 Table 1. SOC mineralization rate ( $\mu$  mol  $m^{-2} s^{-1}$ ) in summer (11 July 2008) and winter (18  
481 November 2008). Data are represented as mean  $\pm$  S.D of five collars.

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

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

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

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

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

Years	F-T				F- $\theta$			
	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	

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538 Table 4. Annual cumulative SOC mineralization rate ( $\text{g C m}^{-2} \text{ year}^{-1}$ ) estimated by linear  
 539 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	$255 \pm 32$	$269 \pm 6$	$374 \pm 65$

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

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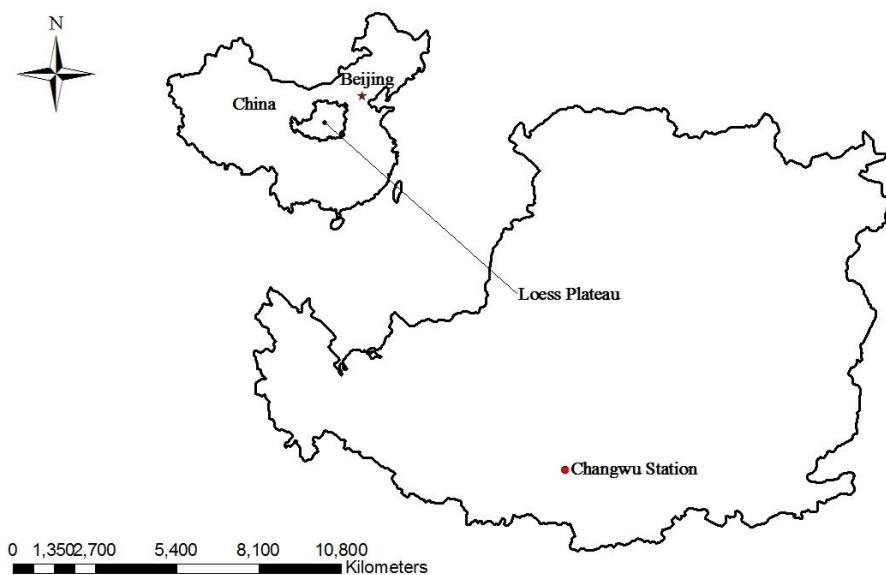
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556 Figure 1. Location of the State Key Agro-Ecological Experimental Station (Changwu  
557 Station).

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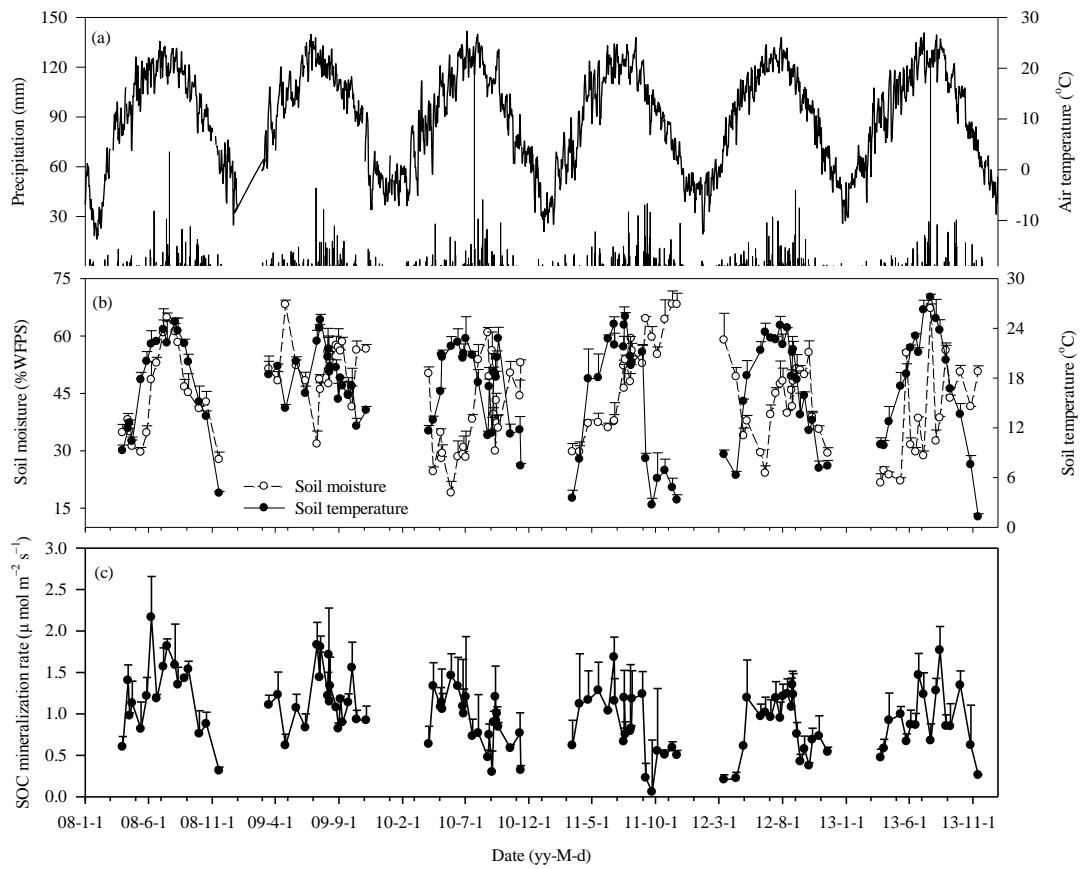
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569 Figure 2. Temporal variations of (a) precipitation and air temperature, (b)  
 570 soil moisture and soil temperature, and (c) SOC mineralization rate from 2008 to 2013.

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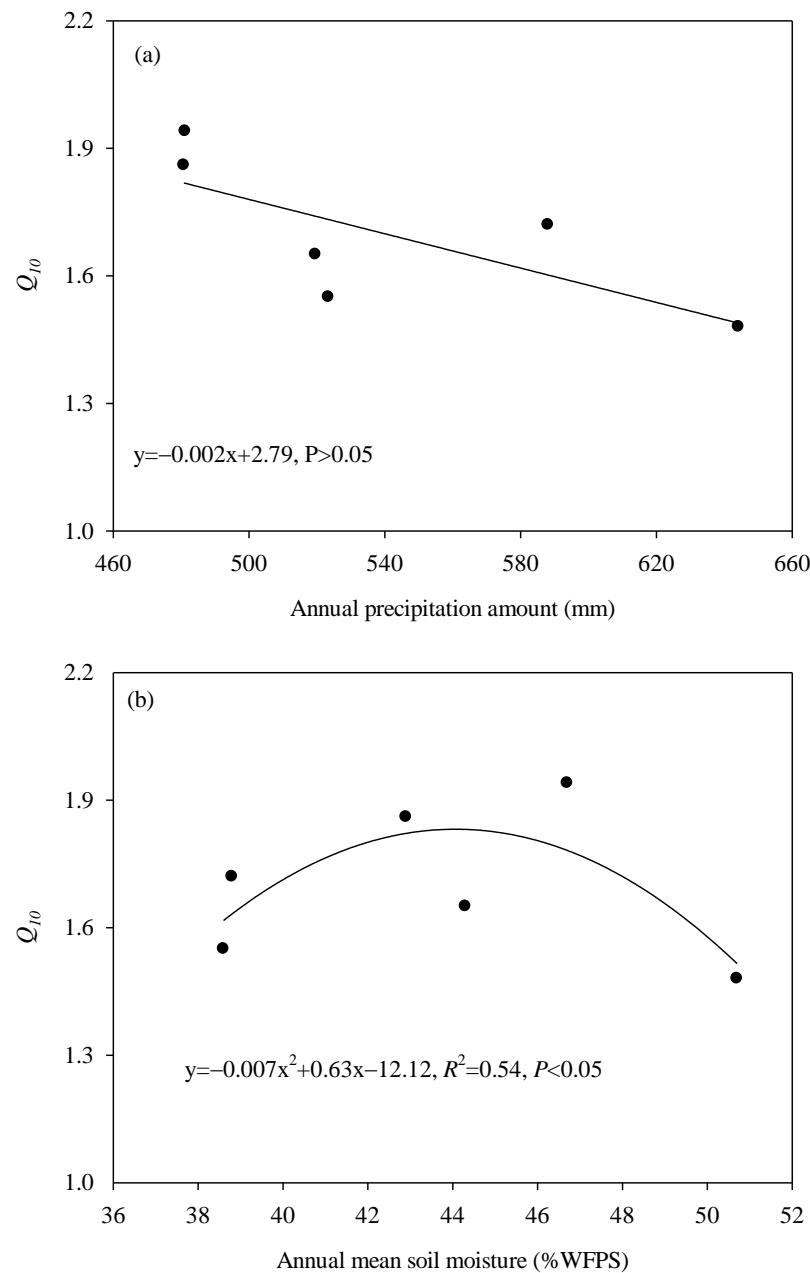
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584 Figure 3. Regression analysis performed between (a)  $Q_{10}$  and annual precipitation amount,  
585 and (b)  $Q_{10}$  and annual mean soil moisture.

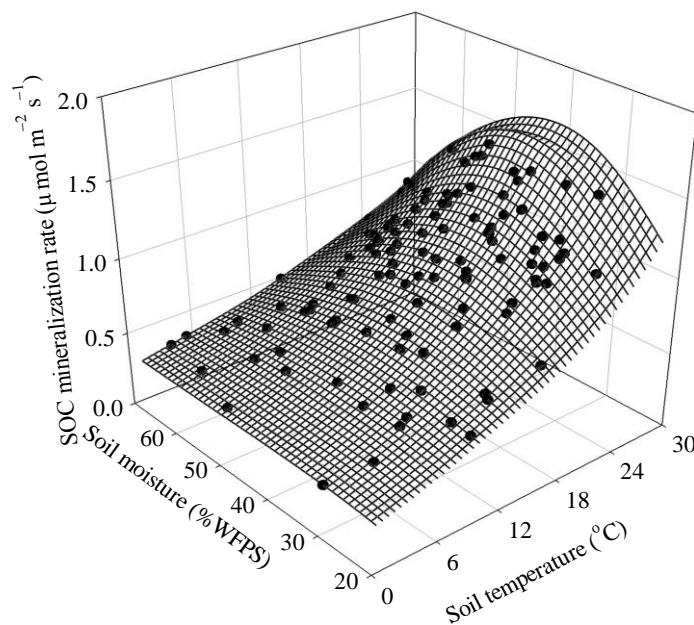
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592 Figure 4. Response surface of SOC mineralization rate as a function of soil moisture and soil  
593 temperature from 2008 to 2013.

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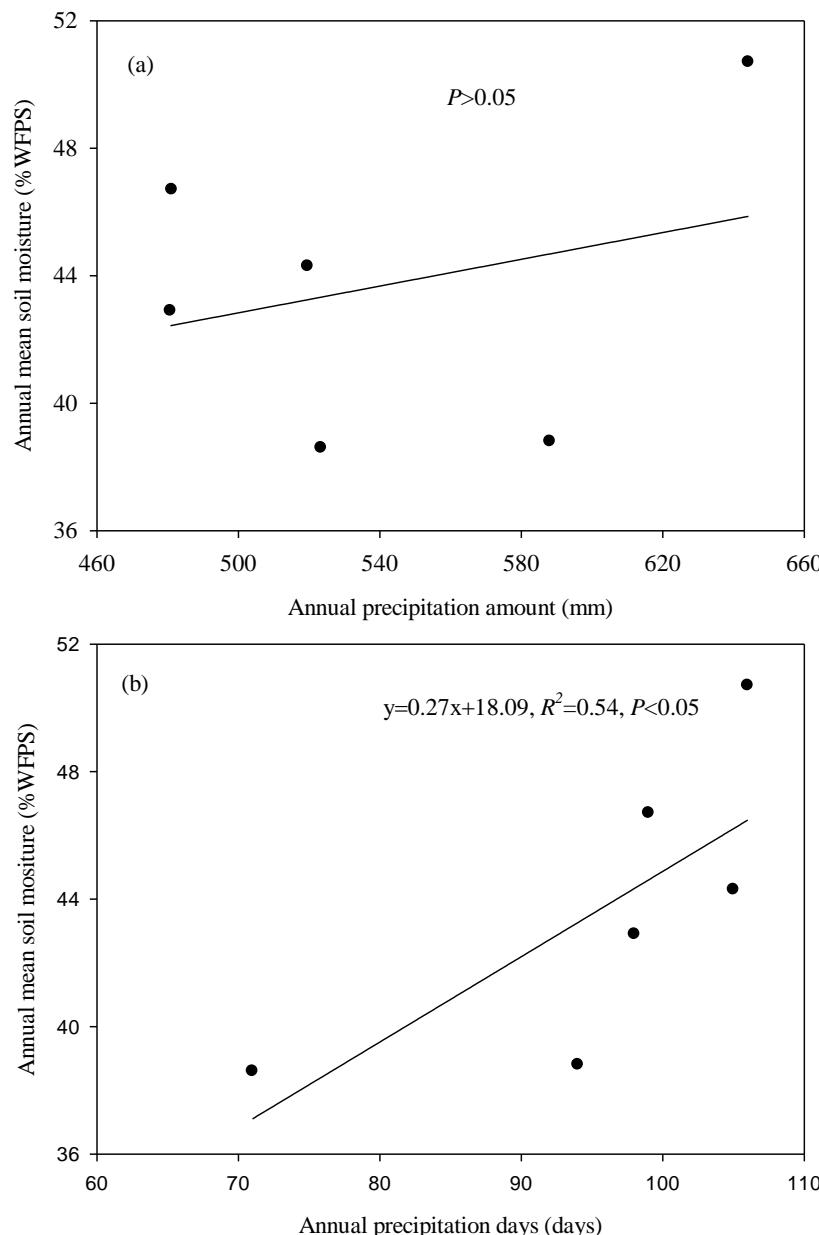
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608 Figure 5. Regression analysis performed between (a) annual mean soil moisture and annual  
609 precipitation amount, and (b) annual mean soil moisture and annual precipitation days.

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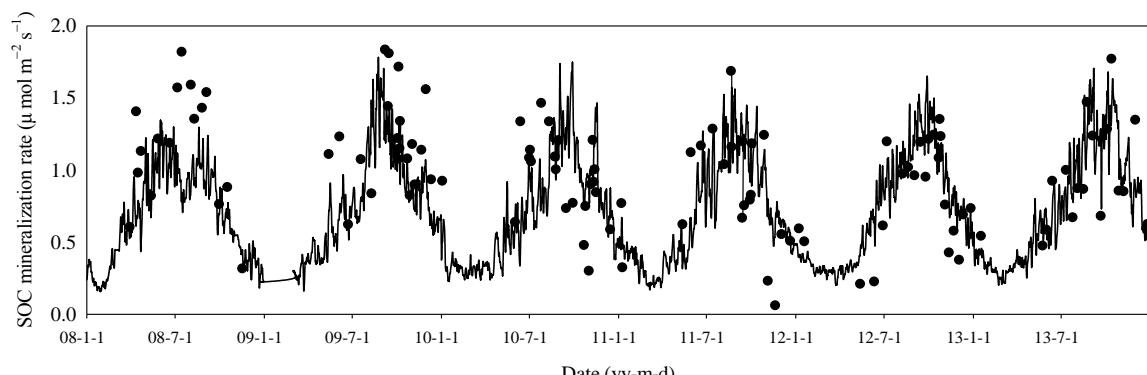
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617 Figure 6. Estimated daily (2008–2013) SOC mineralization rate (solid line) with periodic  
618 measurement values (filled circles).

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