Carbon and Nitrogen Properties of Permafrost over the Eboling Mountain in the Upper Reach of Heihe River Basin, Northwestern China

Authors: Mu, Cuicui, Zhang, Tingjun, Wu, Qingbai, Cao, Bin, Zhang, Xiankai, et al.

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Carbon and nitrogen properties of permafrost over the Eboling Mountain in the upper reach of Heihe River basin, Northwestern China

Cuicui Mu[1]
Tingjun Zhang[1,3]
Qingbai Wu[2]
Bin Cao[1]
Xiankai Zhang[1]
Xiaoqing Peng[1]
Xudong Wan[1]
Lei Zheng[1]
Qingfeng Wang[2] *and*
Guodong Cheng[2]

[1]MOE Key Laboratory of Western China's Environmental Systems, Lanzhou University, Gansu Province 730000, China
[2]State Key Laboratory of Frozen Soil Engineering, Cold and Arid Regions Environmental and Engineering Research Institute, Chinese Academy of Sciences, Lanzhou University, Gansu 730000, China
[3]Corresponding author: College of Earth and Environmental Sciences, Lanzhou University, Gansu 730000, China, tjzhang@lzu.edu.cn

Abstract

The sensitivity of soil carbon and nitrogen to warming is a major uncertainty in projections of climate. However, previous studies about soil organic carbon (SOC) stocks and potential emission predominantly concentrated on the shallow soil layer in high latitude ecosystems. In this study, we analyzed the SOC, total nitrogen (TN) and soil inorganic carbon (SIC) stocks, C/N ratios, and stable carbon isotope ($\delta^{13}C$) in the active layer and permafrost layer on the Eboling Mountain in the upper reach of Heihe River basin, northwestern China. Our results showed that the average stocks of SOC, TN, and SIC in permafrost layer above soil parent materials (71.7 kg m^{-2}, 8.0 kg m^{-2}, 34.7 kg m^{-2}) were much higher than those in the active layer (44.3 kg m^{-2}, 5.3 kg m^{-2}, 12.2 kg m^{-2}). The $\delta^{13}C$ pattern in the soil profiles indicated that historical drainage conditions and pedogenesis were important factors in determining soil organic matter (SOM) stocks in this permafrost region. The $\delta^{13}C$ and C/N ratios of the transient layer and some layers of permafrost implied that the degradation of SOM was different. These results highlight that carbon and nitrogen in permafrost regions with Alpine Kobresia meadow could make significant contribution to China's terrestrial carbon cycle.

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Introduction

Recently, terrestrial ecosystem carbon and nitrogen have received increased attention worldwide because emissions of greenhouse gases into the atmosphere play an important role in driving global warming (Jobbágy and Jackson, 2000; Wynn et al., 2006; Fu et al., 2010). The northern permafrost regions contain approximately 1672 Pg C, which is roughly double the amount of carbon currently in the atmosphere (Tarnocai et al., 2009). It had been proposed that decrease of soil organic carbon (SOC) in permafrost can lead to an increase of greenhouse gases input to the atmosphere (Schuur et al., 2009; Dorrepaal et al., 2009; Smith and Fang, 2010; Schaefer et al., 2011). However, most studies of SOC stocks and its potential emission have focused on soils and upper permafrost (Ping et al., 2008a, 2008b; Zimov et al., 2006), while the deep permafrost carbon (>3 m) remains largely unknown. Especially, most distribution of SOC was studied in high latitude ecosystems, however, magnitude and spatial patterns of carbon and nitrogen in the mid-latitude regions remain largely unknown. Therefore, accurate estimates of carbon stocks, distribution, and its potential emission in deep permafrost of the mid-latitude ecosystems are critical for predicting feedbacks between soil carbon and global warming.

As the largest carbon pool in the terrestrial biosphere, soil carbon stocks consist of organic and inorganic components. However, previous studies dominantly concentrated on SOC stocks, and little is known about the magnitude and patterns of soil inorganic carbon (SIC). The SIC stocks in the top 1 m of soil depth are about 930–1738 Pg (Schlesinger, 2002; Wang et al., 2012). SIC stocks in the Tibetan alpine grasslands are about 2.1 times the corresponding SOC stocks (Yang et al., 2010). Total SIC stocks in China are approximately 60 Pg C, representing 1/20 of the global SIC pool (Wu et al., 2009). The SIC was mostly stored below the 1 m of soil depth (Mikhailova and Post, 2006). Small changes in SOC and SIC may radically alter the carbon balance (Li et al., 2007; Gleixner et al., 2009). SOC and SIC can interact with each other—for example, SOC accumulation with carbonates may induce the dissolution of SIC (Duan et al., 1999)—while respiratory CO_2 from SOM decomposition and root respiration can be used in secondary SIC formation (Entry et al., 2004). Therefore, a concurrent study on the changes in the SIC and SOC in permafrost is important for carbon balance study.

Not only are carbon stocks affected by permafrost degradation, but also soil nitrogen content can play an important role in

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permafrost carbon cycle (Harden et al., 2012). The rate of SOC turnover is strongly linked to nitrogen availability. However, little work has been conducted on nitrogen stocks within the soil profile in comparison to carbon stocks (Crowe et al., 2004). It has long been recognized that nitrogen limitations often constrain carbon accumulations in high latitude ecosystems (Tamm et al., 1982). High nitrogen availability may also strongly increase microbial growth, thus the rate of SOM decomposition (Sistla et al., 2012). Several modeling studies and empirical experiments have identified nitrogen as a potential regulator of SOM decomposition (Churchland et al., 2010; Lavoie et al., 2011). Vertical distribution of nitrogen contents and C/N ratios at different depth in permafrost plays a key role in how soils respond to environmental changes (Callesen et al., 2007). Soil C/N ratios usually are used as a scalar parameter to predict nitrous oxide and methane emissions (Andersson et al., 2012).

Stable carbon isotope analyses of peat and other geological materials is an important tool for gaining insight into biogeochemical processes involved in SOM degradation and preservation (Kracht and Gleixner, 2000), and also in paleoclimate studies (Jones et al., 2010). Detailed sampling of the entire soil profile is crucial to any of these interpretations because $\delta^{13}C$ depth profiles reflect systematic fractionation processes of SOC (Becker-Heidmann and Scharpenseel, 1986).

The Eboling Mountain is located in the Qilian Mountains, northwestern China, and it has experienced severe degradation in recent years. During the International Polar Year (ca. 2007–2008), mean annual ground temperature at depth of 6 m ranged from −3.2 °C to 0.2 °C (Zhao et al., 2010). Few studies about carbon in permafrost were carried out in the alpine permafrost area in the middle latitudes (Wu et al., 2012). Thus it is important to study the deep permafrost here. The aim of this study is to understand the characteristics of soil carbon and nitrogen, especially the dynamics of SOC and SIC in deep permafrost of mid-latitude ecosystem by determining the distribution of SOC, total N, C/N ratios, and $\delta^{13}C$ with depth, and their possible relationships with soil properties. It is important in predicting the response of alpine permafrost in the middle latitudes to global warming.

Field Sampling and Laboratory Analyses

SITE DESCRIPTION AND PERMAFROST SAMPLING

Sampling sites were located in the permafrost zone on the Eboling Mountain in the upper reach of Heihe River basin in the Eastern Qilian Mountain, Northwestern China (Site A: 37°59.872′N, 100°54.977′E, 3700 m), (Site B: 38°00.196′N, 100°54.411′E, 3615 m). The slope of Site A was 11° (north by west 5°) and that of Site B was 7° (north by west 20°) (Mu et al., 2013). The physiography was mountains. Geomorphic positions were the top of slope at Site A and gentle slope at Site B, with the microtopography of turf hummocks. The dominant plant is *Kobresia tibetic*. This watershed is characterized by alpine subhumid climate with the mean annual precipitation (MAP) of 433 mm and the mean annual evaporation of 1080 mm (Wang et al., 2008). It should be noted that about 90% of precipitation falls between May and September in the form of rainfall, and the other 10% occurs from mid-September to mid-May as snow (Liu et al., 2005). Mean annual air temperature of Heihe River basin increased by an average of about 0.6–1 °C during the period 1950 to 2006 (from the daily data of 12 weather stations in the upper reach of Heihe River basin shown in Fig. 1) (Peng et al., 2013). Mean annual ground temperature at depth of 6 m at Site A and B were approximately –0.5 °C and –0.8 °C. Wetlands are prevalent in this study area due to the poor drainage caused by permafrost, and typical periglacial phenomena such as thaw slumping and thermokarst depression are common (Fig. 2) (Mu et al., 2013).

The two permafrost cores were collected by drilling on the north slope of Eboling Mountain from 1 to 14 March 2012. Site A was sampled to depth of 20.4 m and Site B to 11.7 m. Each core was cut into smaller segments, then photographed, wrapped, labeled, and stored in a freezer in the truck. Upon returning to the laboratory of Lanzhou University, the samples were transferred into the ultra-low temperature freezer.

LABORATORY ANALYSES

The samples were stored in the freezer at –50 °C in order to keep the samples in frozen state before processing. Then we conducted samples cutting and analyses of soil physical and chemical characteristics. The subsampling depths for Sites A and B were 600 cm and 470 cm, respectively.

Sample Cutting

All permafrost cores were cut in half lengthwise. Half of each core was cut into about 3–5 cm slices. Every fifth sample was split in duplicate for quality assurance and quality control.

Total Water Content (TWC) and Soil pH

TWC was determined by drying soils at 60 °C for about 48 h and measuring the weight before and after drying (Ahuja et al., 1985). It is expressed as the ratio of the volume of wet soil and the loss mass. The pH of the soil suspension (1:5 soil-water ratio) was measured by an acidity meter (PHS-3C, Shanghai, China).

Bulk Density (BD)

BD was determined by measuring the volume (length, width, height) of a section of frozen core, and then drying the segment at 60 °C and determining its mass.

SOC, SIC, and Elemental Analysis of C and N

SOC, C, and N of homogenized samples were quantified by dry combustion on a Vario EL elemental analyzer (Elemental, Hanau, Germany). For SOC measurement, 0.5 g dry soil samples were pretreated by HCl (10 mL 1 mol L^{-1}) for 24 h to remove carbonate. Then the samples were rinsed with deionized water (millipore) until pH = 7. SIC contents were estimated using the following equation: SIC = Total C – SOC.

Stable Carbon Isotope Analyses

Approximately 2–3 μg of each sample was measured out into tin cups and weighted using an Electronic Precision and Analytical Balance (Sartorius, Germany). Samples were analyzed using OI Analytical Analyzer (Picarro, U.S.A.). Results are based on the mean of three replicates of each sample and expressed as δ-values

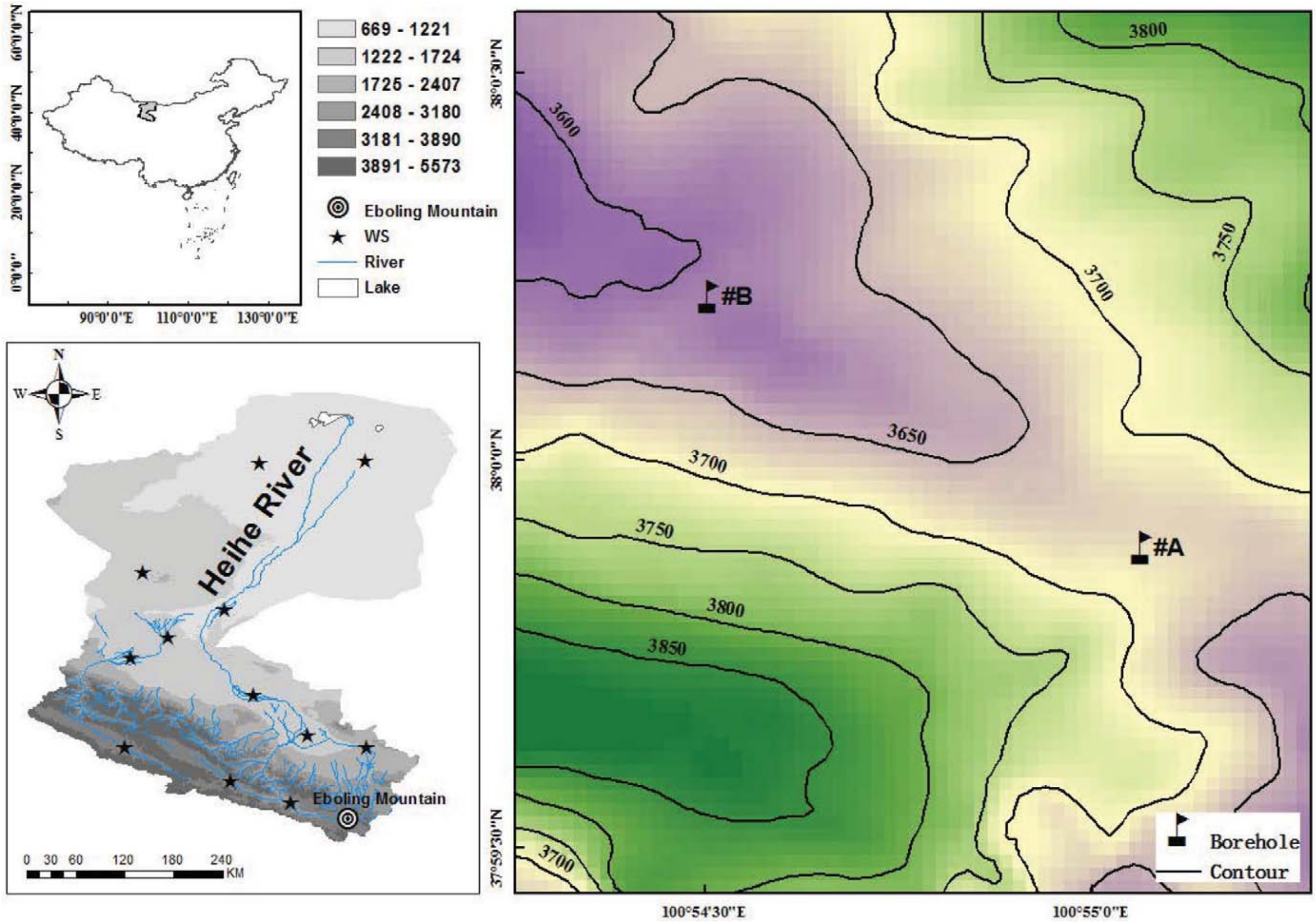

FIGURE 1. Geographic location of the study area and the distribution of sampling sites (stars [WS: weather stations]; flags [sampling sites: Site A and B]). Contour interval = 50 m.

relative to the Vienna Pee Dee Belemnite (VPDB) standard for $\delta^{13}C$. The δ-values are defined as

$$\delta^{13}C = [(R_{sample}/R_{standard}) - 1] \times 1000 \quad (1)$$

where R_{sample} and $R_{standard}$ are $^{13}C/^{12}C$ ratios of the samples and standards, respectively.

CALCULATING SOC, TN, AND SIC STOCKS

The calculation of total SOC stock (SOCS) (kg m^{-2}), soil N stock (TN) (kg m^{-2}), and total SIC stock (SICS) (kg m^{-2}) within soil profile was obtained using Equations 2, 3, and 4.

$$SOCS = \sum_{i=1}^{n} h_i BD_i SOC_i \left(1 - C_i\right) / 100 \quad (2)$$

$$TN = \sum_{i=1}^{n} h_i BD_i N_i \left(1 - C_i\right) / 100 \quad (3)$$

$$SICS = \sum_{i=1}^{n} h_i BD_i SIC_i \left(1 - C_i\right) / 100 \quad (4)$$

where SOC_i = organic carbon (wt%), SIC_i = inorganic carbon (wt%), BD_i = bulk density (g cm^{-3}), N_i = total nitrogen concentration (wt%), h_i = soil thickness (cm), and C_i = percentage of the rock fragments fraction (>2 mm) at layer i, respectively. The C/N ratios were calculated using mass ratios between SOC and TN contents.

Results

SOIL BULK DENSITY, TOTAL WATER CONTENT, AND PH

The general trend for total water content increased with depth gradually with a step increase at depth of about 113 cm of Site A and the depth of about 134 cm of Site B (Figs. 3 and 4). Total water content decreased below the depth of 350 cm at Site B, where a clay layer appeared. There was a good correlation between total water content and bulk density ($P < 0.001$).

Soil bulk density increased with depth and differed at Sites A and B (Figs. 3 and 4). Soil bulk density of Site B was higher

FIGURE 2. Retrogressive thaw slumps and turfs (hummocks) on Eboling Mountain in the upper reaches of Heihe River Basin (eastern branch) (a: turfs [hummocks]; b, c, d: thaw slumps).

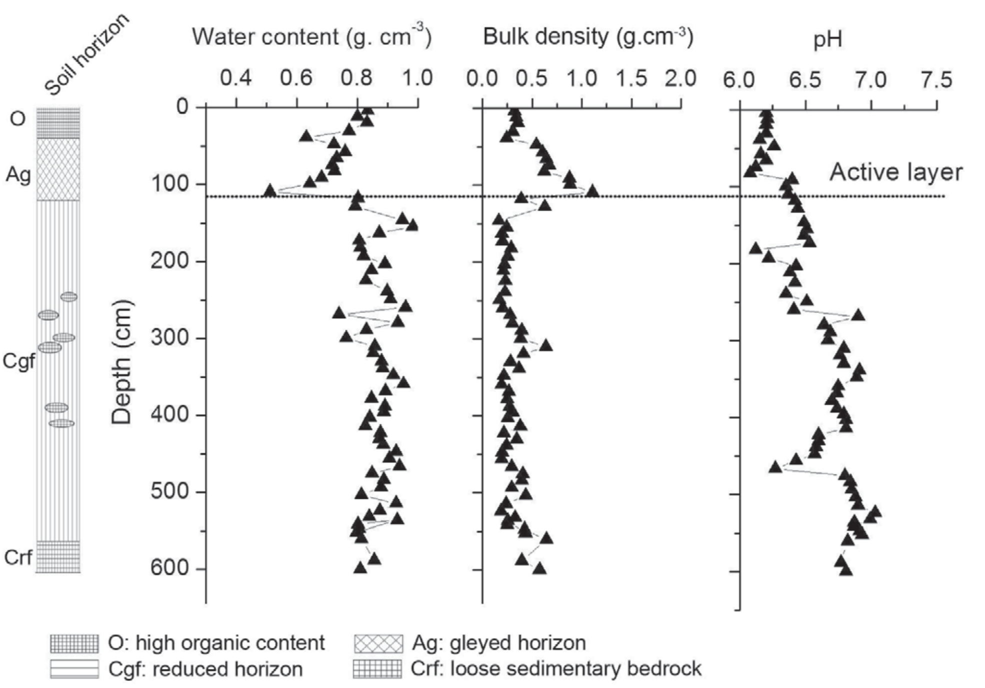

FIGURE 3. Variation of total water content, bulk density, and pH with depth at Site A.

than that of Site A. Soil bulk density had a step increase at the depth from 60 cm to 113 cm of Site A and from 80 cm to 134 cm of Site B. Soil pH values increased with depth at Sites A and B. Soil pH value of Site A ranged from 6.08 to 7.03; the average pH value of the active layer was 6.24, and the average value of the permafrost layer was 6.67. Soil pH value of Site B ranged from 5.95 to 8.46; the average value of the active layer was 6.61, and the average value of the permafrost layer was 7.73.

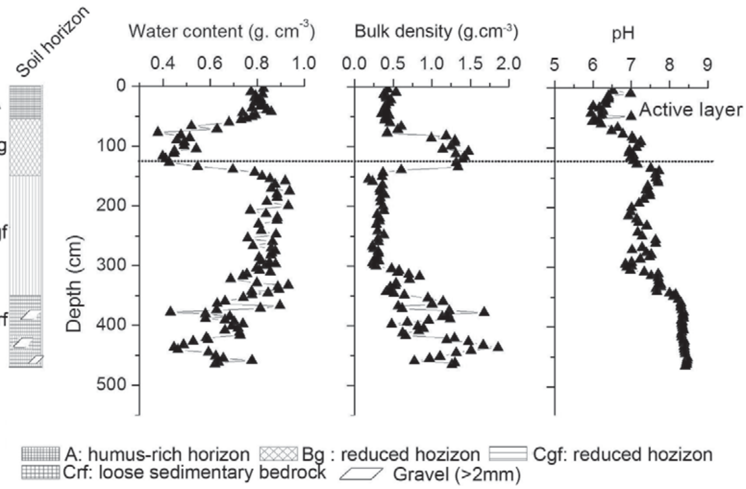

FIGURE 4. Variation of total water content, bulk density, and pH with depth at Site B.

Overall, the pH value in permafrost was higher than that in the active layer at both sites.

SOC, SIC, TN, δ^{13}*C, AND C/N RATIOS*

The SIC concentration of Site A ranged from 0.45% to 12.8%, with an average value of 5.8% (Fig. 5), while the SIC concentration ranged from 0.03% to 17.3% at Site B, with an average value of 4.7% (Fig. 6). SIC decreased with soil depth, and its distribution had the same trend for change with SOC within the soil profile. The average N concentrations were 0.89% and 0.52% at Site A and Site B, respectively. The distribution of TN with depth had the same trend for change with SOC, and the correlation coefficient was 0.95 ($P < 0.001$).

The δ^{13}C values for SOM at Site A ranged from –27.3‰ to –24.9‰, and the average value was –25.8‰ (Fig. 5). The δ^{13}C values below the permafrost table were more negative, ranging from –25.3‰ to –26.7‰, where soil water content was high. The δ^{13}C values of the depth from 350 cm to 460 cm have wider range from –24.9‰ to –27.3‰. The δ^{13}C values for SOM from the top layer to 350 cm at Site B ranged from –28.0‰ to –23.5‰, and the average value was –25.7‰. The δ^{13}C values below the depth of 350 cm were larger than –10.0‰ (Fig. 6).

The soils at Site A had lower C/N ratios and the average value was 7.7 (Fig. 5). The C/N ratios have smaller changes in upper permafrost, reflecting the continuous proportional decreases with depth in the concentrations of SOC and N. The C/N ratios decreased slowly from the active layer to permafrost, then increased toward the bottom of the profile. As for Site B, the C/N ratios ranged from 3.0 to 10.0 within the soil profile (Fig. 6). The C/N ratios below 350 cm at Site B are low because there was excess of inorganic N. The C/N ratios from the depth of 80 cm to 280 cm were higher than those of the depth from 300 cm to 350 cm.

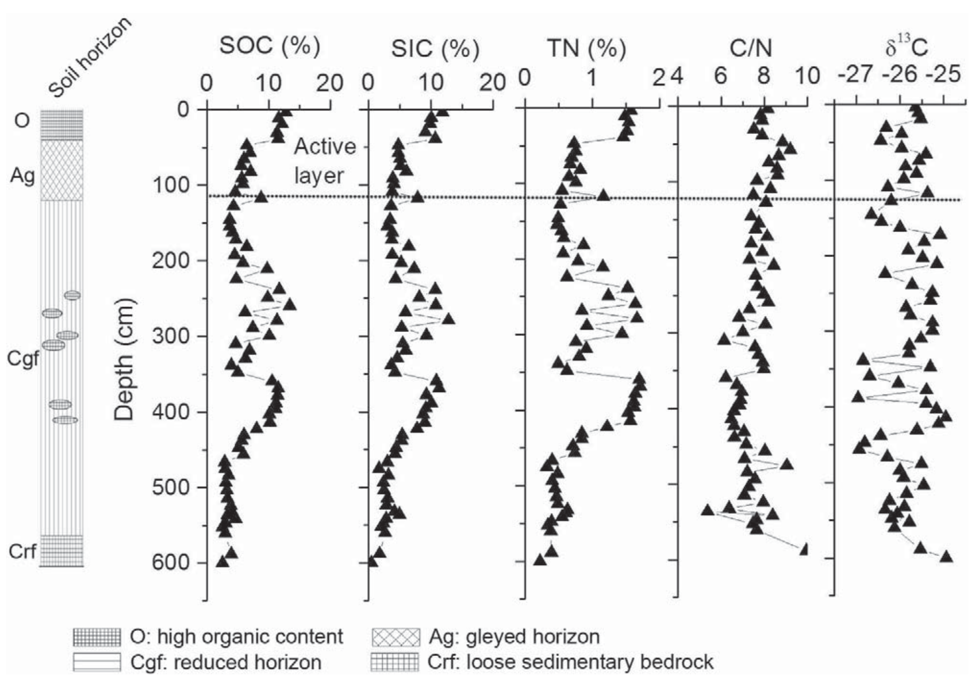

FIGURE 5. Variation of soil organic carbon (SOC), soil inorganic carbon (SIC), total nitrogen (TN), C/N ratios, and δ^{13}C‰ with depth at Site A.

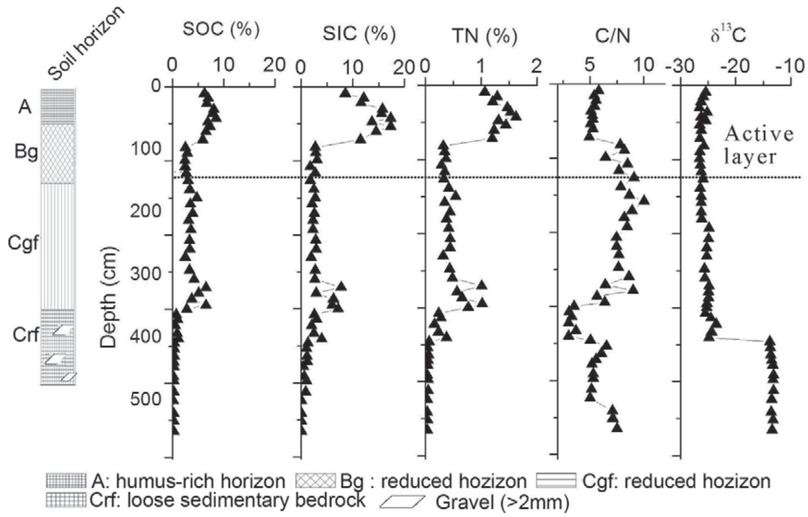

FIGURE 6. Variation of SOC, SIC, TN, C/N ratios, and δ^{13}C‰ with depth at Site B.

SOC, SIC, AND TN STOCKS

The SOC, TN, and SIC stocks in the active layer at Site A were 40.7 kg m^{-2}, 4.9 kg m^{-2}, and 10.1 kg m^{-2}, respectively, and those in permafrost layer were 88.9 kg m^{-2}, 12.2 kg m^{-2}, and 23.4 kg m^{-2}. At Site B, the SOC, TN, and SIC stocks in the active layer were 47.8 kg m^{-2}, 5.7 kg m^{-2}, and 14.2 kg m^{-2}, and those in permafrost layer were 54.5 kg m^{-2}, 3.8 kg m^{-2}, and 45.9 kg m^{-2} (Fig. 7). The average stocks of SOC, TN, and SIC in permafrost layer (71.7 kg m^{-2}, 8.0 kg m^{-2}, 34.7 kg m^{-2}) were probably much higher than those in the active layer (44.3 kg m^{-2}, 5.3 kg m^{-2}, 12.2 kg m^{-2}) because the two permafrost cores could not represent the large area since the pedogenesis was quite complex.

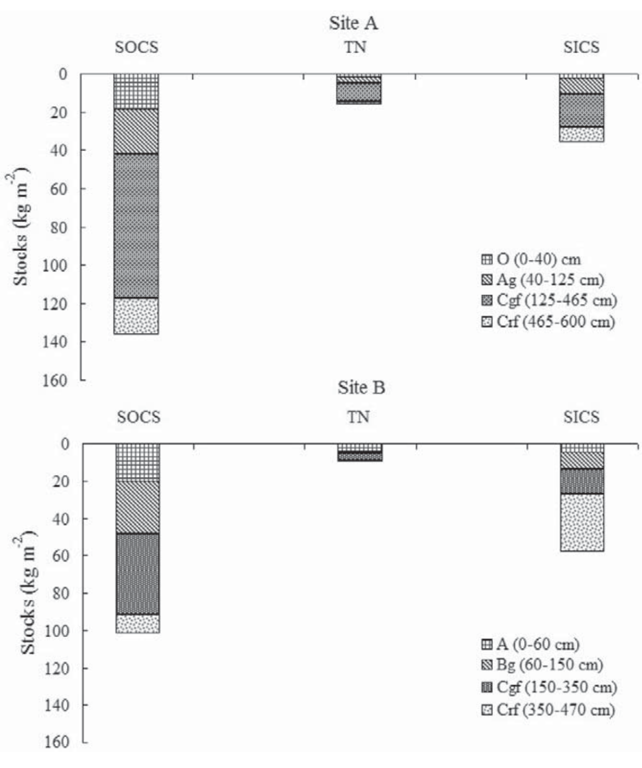

FIGURE 7. Stocks of total SOC (SOCS), total nitrogen (N), and SIC (SICS) at Site A and B within different soil horizons.

Discussion

At the depth of approximately 113 cm of Site A and 134 cm of Site B, ataxitic cryostructure was well developed and total water content was near 1.0 g cm^{-3}. It was reasonable to assume that the transient layer was at ~113–150 cm of Site A and ~134–160 cm of Site B, based on a relatively consistent trend of water content, bulk density, and pH values (Osterkamp et al., 2009; French and Shur, 2010; Murton and French, 1994). There is a wider range in soil bulk density and total water content in the transient layer (Figs. 3 and 4), which is caused by the migration of water in the freeze-thaw process (French, 1999). Studies showed that the transient layer is particularly sensitive to thermokarst due to the high aggradational ice contents (Shur and Jorgenson, 2007; Osterkamp et al., 2009). Thus the thaw slumping and thermokarst depression on the Eboling Mountain are mainly caused by thawing of these ice-rich layers.

The distribution of SOC, N, and SIC have the same trends for all the soil layers at Site B. This result implied that studies of soil carbon balance should include the three components or else an underestimation of the soil carbon sequestration or depletion will occur. The average SIC stocks were 0.6 times of the SOC at the two sites. It is lower than that at depth of 1 m in the Tibetan alpine grasslands, which is about 2.1 times the corresponding SOC stocks (Yang et al., 2010). One explanation is that the moisture of our study area is higher than that of most regions in the Tibetan grassland. High soil moisture can lead to lower SIC content. The extra SIC in the active layer of Site B may come from slope transport and deposits.

High C/N ratios probably reflect better preservation of SOM owing to permafrost or to wetter and anoxic conditions in the peat before the onset of permafrost (Andersson et al., 2012). Soil pH values and C/N ratios had negative correlation in permafrost ($P < 0.001$), which implies alkaline soil environment is favorable to SOM decomposition. The negative relationship between SIC and soil pH ($P < 0.001$) is because SIC can regulate soil acidity (Bronick and Lal, 2005).

Our results support the suggestion by Skrzypek et al. (2010) that δ^{13}C values of SOM were not significantly impacted by diagenesis. The δ^{13}C values were more positive in lower layer, which may indicate that the vascular plants existed here. Fluctuation in δ^{13}C values was greater in the transient layer and the depth from

350 cm to 430 cm relative to the whole profile at Site A, but all $\delta^{13}C$ values were within the range generally associated with C_3 peat-forming plants (Menot and Burns, 2001). The overall trend of $\delta^{13}C$ values displayed a little change, different with the values reported by Oksanen et al. (2001) for Rogovaya River peat plateau in the Russian Arctic, and the values in Andersson et al. (2012) for peat deposit from East European Russian Arctic.

Isotopic depth trends in modern soils that are independent of vegetation changes can be divided into three distinct categories: (1) a $\delta^{13}C$ depth trend toward slightly lower values, (2) a uniform $\delta^{13}C$ depth trend, or (3) a $\delta^{13}C$ depth trend toward higher values (Alewell et al., 2011). The $\delta^{13}C$ values of Site A belonged to the first categories, implying the soils were constantly waterlogged (Stout and Rafter, 1978). The anaerobic conditions apparently produced the small variations in $\delta^{13}C$ values with depth and caused little microbial fractionation of original SOM (Stout et al., 1981). Total water contents, bulk densities, SOC contents, and $\delta^{13}C$ values at Site A can indicate the syngenesis, with the intermittent soil formation due to the aggradation of permafrost caused by periodic addition of new material. The $\delta^{13}C$ values of Site B belonged to the second categories, implying the soils were relatively young and poorly drained soils that underwent more fluctuating, less waterlogged conditions than Site A (Stout and Rafter, 1978). Thus the two sites can represent the different drainage conditions. The uniform $\delta^{13}C$ depth profiles are mainly the result of little time for soil formation, and thus limited decomposable fractionation, which is further impeded by partial waterlogging. Pronounced $\delta^{13}C$ increases are the norm in mature well-drained modern soils or rice land soils (Balesdent and Mariotti, 1996). In addition to aerobic conditions within the soil favoring selective loss of $^{12}CO_2$ accompanying microbial degradation and fractionation (Martin and Haider, 1986), the abundant clays in subsurface were also known to selectively bind ^{13}C (Becker-Heidmann and Scharpenseel, 1986). This is the main reason for higher $\delta^{13}C$ values below the depth of 350 cm at Site B.

The SOC and TN stocks were greater than those of northern China's grasslands (Wu et al., 2003; Yang et al., 2010), and those in Shulehe River basin of Qinghai-Tibetan Plateau, where the average stocks of SOC and TN at depth of 1 m were about 7.72 kg m^{-2} and 0.93 kg m^{-2} (Liu et al., 2012). The SOC and SIC stocks were also greater than those in the region of the *Larix gmelinii* plantations in northeastern China (15.34 kg m^{-2}, 83.38 g m^{-2}) (Wang et al., 2012), and those in the inner Mongolian grasslands (Shi et al., 2012). Though the sampling sites in this study are limited, the two sites can show the potential for distribution and stocks of SOC, TN, and SIC in deep permafrost. This could be because climatic factors and pedogenesis are more favorable to the accumulation of soil carbon and nitrogen in this area (Baumann et al., 2009; Yang et al., 2010).

Conclusions

The transient layers in the permafrost region were at depth of ~113–150 cm of Site A and ~134–160 cm of Site B. The thaw slumping and thermokarst depression on the Eboling Mountain was caused by thawing of the ice-rich layers. The distribution of SIC and TN with depth had the same trend for change with SOC, therefore studies of the soil carbon balance should include three components or else an underestimation of the soil carbon sequestration/depletion will occur. Soil pH value in permafrost layer was higher than that in the active layer. The two sites have different types of drainage based on the isotopic depth trends. The lower $\delta^{13}C$ and C/N values of the transient layer and the depth from 350 cm to 430 cm implied that the SOM was highly degraded at Site A.

The average stocks of SOC, TN, and SIC in permafrost layer above Crf layers (71.7 kg m^{-2}, 8.0 kg m^{-2}, 34.7 kg m^{-2}) were higher than those in the active layer (44.3 kg m^{-2}, 5.3 kg m^{-2}, 12.2 kg m^{-2}). These results highlight that SOC, TN, and SIC in permafrost regions with Alpine Kobresia meadow could make significant contribution to interaction between SOC and SIC in response to climate change in the alpine and plateau of China.

Acknowledgments

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References Cited

Ahuja, L. R., Naney, J. W., and Williams, R. D., 1985: Estimating soil water characteristics from simpler properties of limited data. *Soil Science Society of America Journal*, 49: 1100–1105.

Alewell, C., Giesler, R., Klaminder, J., Leifeld, J., and Rollog, M., 2011: Stable carbon isotopes as indicators for micro-geomorphic changes in palsa peats. *Biogeosciences Discuss*, 8: 527–548.

Andersson, R. A., Meyers, P., Hornibrook, E., Kuhry, P., and Morth, C. M., 2012: Elemental and isotopic carbon and nitrogen records of organic matter accumulation in a Holocene permafrost peat sequence in the East European Russian Arctic. *Journal of Quaternary Science,* 27(6): 545–552.

Balesdent, J., and Mariotti, A., 1996: Measurement of soil organic matter turnover using ^{13}C natural abundance. *In* Boutton, T. W., and Yamasaki, S. (eds.), *Mass Spectrometry of Soils*. New York: Marcel-Dekker, 83–111.

Baumann, F., He, J. S., Schmidt, K., Kuhn, P., and Scholten, T., 2009: Pedogenesis, permafrost, and soil moisture as controlling factors for soil nitrogen and carbon contents across the Tibetan Plateau. *Global Change Biology*, 15: 3001–3017.

Becker-Heidmann, P., and Scharpenseel, H. W., 1986: Thin layer $\delta^{13}C$ and $\delta^{14}C$ monitoring of ''lessive'' soil profiles. *Radiocarbon*, 28: 383–390.

Bronick, C. J., and Lal, R., 2005: Soil structure and management: a review. *Geoderma*, 124: 3–22.

Callesen, I., Raulund–Rasmussen, K., Westman, C. J., and Tau-Strand, L., 2007: Nitrogen pools and C:N ratios in well-drained Nordic forest soils related to climate and soil texture. *Boreal Environment Research*, 12: 681–692.

Churchland, C., Mayo-Bruinsma, L., Ronson, A., and Grogan, P., 2010: Soil microbial and plant community responses to single large carbon and nitrogen additions in Low Arctic tundra. *Plant and Soil*, 334: 409–421.

Crowe, A. M., Sakata, A., Mcclean, C., and Cresser, M. S., 2004: What factors control soil profile nitrogen storage? *Water, Air, and Soil Pollution Focus*, 4: 75–84.

Dorrepaal, E., Toet, S., van Logtestijn, R. S. P., Swart, E., van de Weg, M. J., Callaghan, T. V., and Aerts, R., 2009: Carbon respiration from subsurface peat accelerated by climate warming in the subarctic. *Nature*, 460: 616–620.

Duan, J. N., Li, B. G., Shi, Y. C., Yan, T. L., and Zhu, D. H., 1999: Modeling of soil $CaCO_3$ deposition process in arid areas. *Acta Pedol Sin*, 36: 318–326.

Entry, J. A., Sojka, R. E., and Shewmaker, G. E., 2004: Irrigation increases inorganic carbon in 550 agricultural soils. *Environ Manage*, 33: 309–317.

French, H., 1999: Past and present permafrost as an indicator of climate change. *Polar Research*, 18(2): 269–274.

French, H., and Shur, Y., 2010: The principles of cryostratigraphy. *Earth-Science Reviews*, 101 (3–4): 190–206.

Fu, X. L., Shao, M. G., Wei, X. R., and Robert, H., 2010: Soil organic carbon and total nitrogen as affected by vegetation types in Northern Loess Plateau of China. *Geoderma*, 155: 31–35.

Gleixner, G. D., Tefs, C., Jordan, A., et al., 2009: Soil carbon accumulation in old-growth forests. *In* Wirth, C., Gleixner, G., and Heimann, M. (eds.), *Old-Growth Forests: Function, Fate and Value.* Berlin: Springer, 231–266.

Harden, J. W., Koven, C. D., Ping, C. L., Hugelius, G., McGuire, A. D., Camill, P., and Jorgenson, T., 2012: Field information links permafrost carbon to physical vulnerabilities of thawing. *Geophysical Research Letters*, 39: L15704, http://dx.doi.org/10.1029/2012GL051958.

Jobbágy, E. G., and Jackson, R. B., 2000: The vertical distribution of soil organic carbon and its relation to climate and vegetation. *Ecological Applications,* 10(2): 423–436.

Jones, M. C., Peteet, D. M., and Sambrotto, R., 2010: Late-glacial and Holocene $\delta^{15}N$ and $\delta^{13}C$ variation from a Kenai Peninsula, Alaska peatland. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 293: 132–143.

Murton, J. B., and French, H. M., 1994: Cryostructures in permafrost, Tuktoyaktuk coastlands, western Arctic Canada. *Canadian Journal of Earth Sciences*, 31(4): 737–747.

Kracht, O., and Gleixner, G., 2000: Isotope analysis of pyrolysis products from *Sphagnum* peat and dissolved organic matter from bog water. *Organic Geochemistry*, 31: 645–654.

Lavoie, M., Mack, M. C., and Schuur, E. A. G., 2011: Effects of elevated nitrogen and temperature on carbon and nitrogen dynamics in Alaskan Arctic and boreal soils. *Journal of Geophysical Research,* 116: G03013, http://dx.doi.org/10.1029/2010JG001629.

Li, Z. P., Han, F. X., Su, Y., Zhang, T. L., Sun, B., Monts, D. L., and Plodine, M. J., 2007: Assessment of soil organic and carbonate carbon storage in China. *Geoderma*, 138: 119–126.

Liu, W. J., Chen, S. Y., Qin, X., Baumann, F., Scholten, T., Zhou, Z., Sun, W. J., Zhang, T. Z., Ren, J. W., and Qin, D. H., 2012: Storage, patterns, and control of soil organic carbon and nitrogen in the northeastern margin of the Qinghai–Tibetan Plateau. *Environmental Research Letters,* 7: 8–9.

Liu, X. H., Qin, D. H., Shao, X. M., Chen, T., and Ren, J. W., 2005: Temperature variations recovered from tree-rings in the middle Qilian Mountain over the last millennium. *Earth Science*, 48: 521–529.

Martin, A., and Haider, K., 1986: Influence of mineral colloids on the turnover rates of soil organic carbon. *In* Huang, P. M., and Schnitzer, M. (eds.), *Interactions of Soil Minerals with Natural Organics and Soil Microbes*. Madison, Wisconsin: Soil Science Society of America Special Publication 17, 283–304.

Menot, G., and Burns, S. J., 2001: Carbon isotopes in ombrogenic peat bog plants as climatic indicators: calibration from an altitudinal transect in Switzerland. *Organic Geochemistry*, 32: 233–245.

Mikhailova, E. A., and Post, C. J., 2006: Effects of land use on soil inorganic carbon stocks in the Russian Chernozem. *Journal of Environmental Quality,* 35: 1384–1388.

Mu, C. C., Zhang, T. J., Cao, B., Wan, X. D., Peng, X. Q., and Cheng, G. D., 2013: Study of the organic carbon storage in the active layer of the permafrost over the Eboling Mountain in the upper reaches of the Heihe River in the Eastern Qilian Mountains. *Journal of Glaciology and Geocryology*, 35: 1–9.

Mu, C. C., Zhang, T. J., Wu, Q. B., et al., 2014: Stable carbon isotopes as indicators for permafrost carbon vulnerability in upper reach of Heihe River basin, northwestern China. *Quaternary International*, 321: 71–77.

Oksanen, P. O., Kuhry, P., and Alekseeva, R. N., 2001: Holocene development of the Rogovaya River peat plateau, European Russian Arctic. *The Holocene*, 11: 25–40.

Osterkamp, T. E., Jorgenson, M. T., Schurr, E. A. G., Shur, Y. L., Kanevskiy, M. Z., Vogel, J. G., and Tumskoy, V. E., 2009: Physical and ecological changes associated with warming permafrost and thermokarst in interior Alaska. *Permafrost and Periglacial Processes*, 20: 235–256.

Ping, C. L., Michaelson, G. J., Jorgenson, T., Kimble, J. M., Epstein, H., Romanovsky, V. E., and Walker, D. A., 2008a: High stocks of soil organic carbon in the North American Arctic region. *Nature Geoscience*, 1: 615–619.

Ping, C. L., Michaelson, G. J., Kimble, J. M., Romanovsky, V. E., Shur, Y. L., Swanson, D. K., and Walker, D. A., 2008b: Cryogenesis and soil formation along a bioclimate gradient in Arctic North America. *Journal of Geophysical Research,* 113: G03S12.

Peng, X. Q., Zhang, T., Pan, X., et al., 2013: Spatial and temporal variations of seasonally frozen ground over the Heihe River Basin of Qilian Mountain in western China. *Advances in Earth Science,* 28(4): 497–508.

Schaefer, K., Zhang, T., Bruhwiler, L., and Barrett, A. P., 2011: Amount and timing of permafrost carbon release in response to climate warming. *Tellus B,* 63: 165–180.

Schlesinger, W. H., 2002: Inorganic carbon and the global carbon cycle. *Encyclopedia of Soil Science.* New York: Marcel Dekker.

Schuur, E. A. G., Vogel, J. G., Crummer, K. G., Lee, H., Sickman, J. O., and Osterkamp, T. E., 2009: The effect of permafrost thaw on old carbon release and net carbon exchange from tundra. *Nature,* 459: 556–559.

Shi, Y., Baumann, F., Ma, Y., Song, C., Kuhn, P., Scholten, T., and He, J. S., 2012: Organic and inorganic carbon in the topsoil of the Mongolian and Tibetan grasslands: pattern, control, and implications. *Biogeosciences Discussions*, 9: 1869–1898.

Shur, Y. L., and Jorgenson, M. T., 2007: Patterns of permafrost formation and degradation in relation to climate and ecosystems. *Permafrost and Periglacial Processes*, 18: 7–19.

Sistla, S. A., Asao, S., and Schime, J. P., 2012: Detecting microbial N-limitation in tussock tundra soil: implications for Arctic soil organic carbon cycling. *Soil Biology & Biochemistry,* 55: 78–84.

Skrzypek, G., Jezierski, P., and Szynkiewicz, A., 2010: Preservation of primary stable isotope signatures of peat-forming plants during early decomposition—observation along an altitudinal transect. *Chemical Geology*, 273: 238–249.

Smith, P., and Fang, C. M., 2010: A warm response by soils. *Nature*, 464: 499–500.

Stout, J. D., and Rafter, T. A., 1978: *The $^{13}C/^{12}C$ Isotopic Ratios of Some New Zealand Tussock Grassland Soils: Stable Isotopes in the Earth Sciences*. New Zealand Department of Scientific and Industrial Research, DSIR (Department of Scientific and Industrial Research) Bulletin 220, INS (Institute of Nuclear Sciences) Contribution 806: 75–83.

Stout, J. D., Goh, K. M., and Rafter, T. A., 1981: Chemistry and turnover of naturally occurring resistant organic compounds in soil. *In* Paul, E. A., and Ladd, J. N. (eds.), *Soil Biochemistry*. New York: Marcel Dekker, 19–24.

Tamm, C. O., Lake, J. V., and Miller, H. G., 1982: Nitrogen cycling in undisturbed and manipulated boreal forest. *Philosophical Transactions of the Royal Society B*, 296: 419–425.

Tarnocai, C., Canadell, J. G., Schuur, E. A. G., Kuhry, P., Mazhitova, G., and Zimov, S., 2009: Soil organic carbon pools in the northern circumpolar permafrost region. *Global Biogeochemical Cycles,* 23: 1–11.

Wang, J. Y., Yu, P. T., and Wang, Y. H., 2008: *Ecohydrological Processes in Forests: Case Study from Qilian Mountains*. Beijing: Science Press.

Wang, W. J., Su, D. X., Qiu, L., Wang, H. Y., An, J., Zheng, G. Y., and Zu, Y. G., 2012: Concurrent changes in soil inorganic and organic carbon during the development of larch, *Larix gmelinii,* plantations and their effects on soil physicochemical properties. *Environmental Earth Sciences,* 69: 1559–1570.

Wu, H. B., Guo, Z. T., and Peng, C. H., 2003: Distribution and storage of soil organic carbon in China. *Global Biogeochemical Cycles,* 17: 1048.

Wu, H., Guo, Z., Gao, Q., and Peng, C., 2009: Distribution of soil inorganic carbon storage and its changes due to agricultural land use activity in China. *Agriculture, Ecosystems & Environment*, 129: 413–421.

Wu, X. D., L. Zhao, H. B. Fang, G. Y. Yue, J. Chen, Q. Q. Pang, Z. W. Wang, and Y. J. Ding, 2012: Soil organic carbon and its relationship to vegetation communities and soil properties in permafrost of middle-western Qinghai-Tibet Plateau. *Permafrost and Periglacial Process*, 23: 162–169.

Wynn, J. G., Bird, M. I., Vallen, L., Grand-Clement, E., Carter, J., and Berry, S. L., 2006: Continental-scale measurement of the soil organic carbon pool with climatic, edaphic, and biotic controls. *Global Biogeochemical Cycles*, 20(1): GB1007, http://dx.doi.org/10.1029/2005GB002576.

Yang, Y. H., Fang, J. Y., Tang, Y. H., Ji, J. S., and Zhu, B., 2008: Storage, patterns and controls of soil organic carbon in the Tibetan grasslands. *Global Change Biology,* 14: 1592–1599.

Yang, Y. H., Fang, J. Y., Ji, C. J., Ma, W. H., Su, S. S., and Tang, Z. Y., 2010: Soil inorganic carbon stock in the Tibetan alpine grasslands. *Global Biogeochemical Cycles,* 24: http://dx.doi.org/10.1029/2010GB003804.

Zhao, L., Wu, Q., and Marchenko, S., et al., 2010: Thermal state of permafrost and active layer in Central Asia during the international polar year. *Permafrost and Periglacial Processes*, 21: 198–207.

Zimov, S. A., Schuur, E. A. G., and Chapin, F. S., 2006: Permafrost and the global carbon budget. *Science,* 312: 1612–1613.

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