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Organic and inorganic carbon in the topsoil of the Mongolian and Tibetan grasslands: pattern, control and implications

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Abstract

Soil carbon (C) is the largest C pool in terrestrial biosphere and includes both inorganic and organic components. Studying patterns and controls of soil C help us to understand and estimate potential responses of soil C to global change in the future. Here we analyzed topsoil data of 81 sites obtained from a regional survey across grasslands in the Inner Mongolia and on the Tibetan Plateau during 2006–2007, attempting to find the patterns and controls of soil inorganic carbon (SIC) and soil organic carbon (SOC). The average of SIC and SOC in the topsoil (0–20 cm) across the study region were 0.38 % and 3.63 %, ranging between 0.00–2.92 % and 0.32–26.17 %, respectively. Both SIC and SOC in the topsoil of the Tibetan grasslands (0.51 % and 5.24 %, respectively) were higher than those of the Inner Mongolian grasslands (0.21 % and 1.61 %). Regression tree analyses showed that the spatial pattern of SIC and SOC were controlled by different factors. Chemical and physical processes of soil formation drive the spatial pattern of SIC, while biotic processes drive the spatial pattern of SOC. SIC was controlled by soil acidification and other processes depending on soil pH. Vegetation type is the most important variable driving the spatial pattern of SOC. According to our models, given the acidification rate in Chinese grassland soils in the future is the same as that in Chinese cropland soils during the past two decades: 0.27 and 0.48 units per 20 yr in the Inner Mongolian grasslands and the Tibetan grasslands, respectively, it will lead to 30 % and 53 % decrease in SIC in the Inner Mongolian grasslands and the Tibetan grasslands, respectively. However, negative relationship between soil pH and SOC suggests that acidification will inhibit decomposition of SOC, thus will not lead to a significant general loss of carbon from soils in these regions.

1 Introduction

The importance of soil carbon (C) in global C cycling has received considerable attention in recent decades (Schulze and Freibauer, 2005; Trumbore and Czimczik, 2008;

Schlesinger and Andrews, 2000; Lal, 2004a). Soil C pool comprises two distinct components: soil organic carbon (SOC) and soil inorganic carbon (SIC). The latter includes lithogenic inorganic carbon (LIC), which comes from parent material, and pedogenic inorganic carbon (PIC), which is formed through the dissolution and precipitation of carbonate parent material. On a global scale, the estimated size of SOC in the 1-m depth is 1500 to 1600 Pg (Lal, 2004b; Monger and Gallegos, 2000; Batjes, 1996), and the size of SIC is 695 to 1738 Pg (Batjes, 1996; Eswaran et al., 1995).

C flux between the atmosphere and terrestrial ecosystems involves both SOC and SIC. The C flux between SOC pools and the atmosphere depends on biomass production, organic matter input and soil respiration (Schlesinger and Andrews, 2000; Valentini et al., 2000). SIC pools exchange C with the atmosphere through a series of physical and chemical reaction, such as C sequestration by carbonate formation or CO_2 release by acidification and leaching (Lal and Kimble, 2000; Lal, 2008; Ouyang et al., 2008). Given that global change has altered temperature, precipitation, nitrogen availability and many other environmental factors (Rockström et al., 2009; Vitousek, 1994), these changes are most likely to have great impact on soil carbon. As the largest C pool in terrestrial biosphere, even a minor change in soil carbon stocks could result in a significant alteration of atmospheric CO_2 concentration (Davidson and Janssens, 2006; Trumbore and Czimczik, 2008). Therefore, both SIC and SOC pools should be considered in order to more accurately predict future soil carbon dynamics.

In arid and semiarid regions, which cover as much as one-third of the earth's surface, SIC pools and their dynamics are important as the rate of accumulation of SIC is generally higher than in other biomes (Lal, 2008). Grasslands are one of the most widespread ecosystems in arid and semiarid regions, containing about 20 % of global soil C stocks (Jobbagy and Jackson, 2000; Scurlock and Hall, 1998; Wang and Fang, 2009). Grasslands cover more than 40 % of China's land surface, ranging from temperate grasslands in arid/semi-arid region to alpine grasslands on the Tibetan Plateau (Kang et al., 2007; Yang et al., 2010b; He et al., 2009). Because of the large spatial extend, Chinese grasslands have significant effects on regional and global carbon cycles

(Ni, 2002; Yang et al., 2010b). Moreover, Chinese grasslands face rapid environmental changes (Kang et al., 2007). In recent years, many studies have been conducted to investigate the pattern, controls and dynamic of SOC (e.g., Ni, 2002; Yang et al., 2010b; Wu et al., 2003; Yu et al., 2007; Xie et al., 2007; Baumann et al., 2009) as well as SIC (e.g., Yang et al., 2010a; Mi et al., 2008; Feng et al., 2002) in Chinese grasslands. SOC stocks are reported to range from 37.7 Pg (Xie et al., 2007) to 41.0 Pg (Ni, 2002) in Chinese grasslands, and about 16.7 Pg in Northern China's grasslands (Yang et al., 2010b). On the scale of whole China, SOC densities are significantly influenced by temperature (Xie et al., 2007), while soil moisture and texture control SOC density in the Tibetan grasslands (Yang et al., 2008; Baumann et al., 2009). For SIC, total stocks of the top 1 m are estimated to vary from 53.3 Pg to 77.9 Pg (Wu et al., 2009; Mi et al., 2008; Li et al., 2007) in China. Yang et al. (2010a) estimated the SIC stock of the Tibetan grasslands to 15.2 Pg. Temperature and precipitation both show significant relationships with SIC density (Li et al., 2007; Mi et al., 2008; Yang et al., 2010a).

However, there are still some uncertainties evident in recent studies. Most of the researches focused on mean soil carbon stocks for soil at depth of 1 m or more, but little work, especially for SIC, has been devoted to investigate the patterns, controls and dynamics of soil carbon in the topsoil. The topsoil is the component of the soil system showing most rapid responses to environmental changes, such as alterations in temperature, precipitation and nitrogen deposition (Liao et al., 2009; Song et al., 2005). Changes in the topsoil are particularly important for exploring ecosystem response and functioning (Franzluebbers and Stuedemann, 2010). Moreover, recent studies spent large efforts in determining pattern, controls and dynamic of SOC, but there is less attention drawn on SIC in general (Eswaran et al., 2000; Mi et al., 2008). Consequently, we cannot simply use results of these studies to predict potential responses of soil carbon to future global changes. Further, some recent studies calculated SIC as the difference between soil total carbon (STC) and SOC (e.g., Yang et al., 2010a; Mikhailova and Post, 2006). However, the proportions of carbonates in soil total carbon are usually small (Chatterjee et al., 2009), thus this method without the direct measurement of SIC could produce a large relative error. Therefore, studies focused on SIC based on measured data help us to reduce the uncertainties of previous studies and to predict response of soil carbon to global changes.

In this study, we analyzed the topsoil data of 81 sites obtained from a regional survey across grasslands in Inner Mongolia and the Tibetan Plateau during 2006–2007. We attempt (1) to investigate the large-scale patterns of SIC and SOC in the topsoil of grasslands in Inner Mongolia and on the Tibetan Plateau, (2) to examine the changes of SIC and SOC along naturally occurring environmental gradients in order to identify key controlling factors for SIC and SOC contents, and (3) to estimate potential responses of SIC and SOC in the topsoil related to soil acidification.

2 Materials and methods

2.1 Study sites

This study was conducted in temperate grasslands in the Inner Mongolia and alpine grasslands on the Tibetan Plateau, during two expeditions in late July and early August of 2006 and 2007. We selected 81 sites along an approximately 4000 km long transect (longitude from 90.80 to 120.12° E, latitude from 30.31 to 50.19° N, and altitude from 549 to 5105 m a.s.l.) for soil and plant community sampling (Table 1, Fig. 1). Mean annual temperature (MAT) ranges from −5.8 to 4.1 °C and mean annual precipitation (MAP) ranges from 148 to 604 mm. The sites along the transect represent natural zonal grassland vegetation, including the five main vegetation types: meadow steppe, typical steppe, desert steppe, alpine steppe and alpine meadow. Field sites were selected by visual inspection of the vegetation, aiming to select sampling sites subject to minimal grazing and other anthropogenic disturbances.

2.2 Soil and biomass survey

Detailed field investigations included soil profile description according to FAO (2006) and IUSS Working Group WRB (2006). Soil sampling was split into two parts: schematic soil sampling by drilling at three depth-increments (0–5, 5–10, and 10–20 cm) for SIC and SOC analyses, as well as volumetric sampling using a standard container (100 cm^3 in volume) at equal depths for soil bulk density (SBD) and gravimetric water content (equalling to soil moisture, SM) determination. The sampling protocol is described in Baumann et al. (2009).

Soil samples were air-dried, all live root material was removed and the remaining soil was ground using a ball mill (NM200, Retsch, Germany). We measured soil inorganic carbon (SIC) volumetrically using an inorganic carbon analyzer (Calcimeter 08.53, Eijkelkamp, Netherland). Soil total carbon (STC) was measured by dry combustion using an elemental analyzer (VARIO EL III, Elementar, Hanau, Germany) with a combustion temperature of 1150 °C and a reduction temperature of 850 °C. Soil organic carbon (SOC) was calculated as the difference between STC and SIC. The average concentration of SIC, SOC and STC of the top 20 cm was calculated using the data from the three layers with soil bulk density, and this average value was used as site level data. Soil pH was determined in 0.01 M $CaCl_2$ potentiometrically. Soil total nitrogen was also determined using an elemental analyzer (PE 2400 II CHN elemental analyzer, Perkin-Elmer, Boston, Massachusetts, USA).

Additionally, we measured aboveground biomass (AGB) in three plots ($1 \times 1\,m^2$) and belowground biomass (BGB) in three soil pits ($0.5 \times 0.5\,m^2$) at each site. Biomass was measured by oven-drying at 60 °C to a constant weight and weighting to the nearest 0.1 g.

2.3 Climate data and statistical analyses

Climate data for each site was calculated based on linear models using latitude, longitude and altitude as predictors from 55-yr average temperature and precipitation records (1951–2005) at 680 climate stations across China (Fang et al., 2001; He et al., 2009). The calculation of potential evapotranspiration (PE) and actual evapotranspiration (AE) were based on Thornthwaite's method (Thornthwaite, 1948).

We used one-way ANOVA with Turkey's *post hoc* test to compare the effects of region, vegetation type, and soil depth on SIC, SOC and the ratio of SIC to STC (SIC/STC). We conducted classification and regression tree (CART) analyses to detect important variables influencing the patterns of SIC, SOC and SIC/STC. We selected the complexity parameter as split criterion, and set the observations required for a split search at 15. Climatic factors, including MAT, MAP, growing season temperature (GST, from April to August), growing season precipitation (GSP), AE, PE, as well as biotic factors including vegetation type (VT), AGB, BGB, were used in the tree models (Table 4). Since soil pH, SM, CO_2 partial pressure and water deficiency also influenced SIC, soil pH, SM, altitude (proxy of CO_2 partial pressure) as well as the ratio of MAP to PE (MAP/PE, an index of water deficiency) were incorporated into the CART analysis when we analysed SIC and SIC/STC. Finally, for estimating the change of SIC of the topsoil in the Inner Mongolian and Tibetan grasslands under the background of soil acidification, we built ordinary linear regression models for SIC and SIC/STC, using the most powerful explanatory variables in the CART analysis. SIC, SOC and SIC/STC were log transformation to achieve normal distribution (Fig. 2). All statistical analyses were performed using R 2.3.0 (R Development Core Team, 2011). The classification and regression trees were developed using the R package rpart.

3 Results

3.1 Overall patterns of SIC, SOC and SIC/STC

Frequency distributions of SIC, SOC and SIC/STC did not deviate significantly from log normal distributions ($P > 0.05$) neither at site level nor for each depth increment. At site level, the mean values of SIC, SOC and SIC/STC were 0.38 %, 3.58 % and 0.15,

ranging between 0.00–2.92 %, 0.32–26.34 % and 0.00–0.65, with CV of 1.36, 1.19 and 1.26, respectively (Table 3). We did not find significant differences of SIC and SIC/STC among three soil depths ($P < 0.05$), while SOC in 0–5 cm was significantly higher than that in 10–20 cm.

At site level, topsoil SIC and SOC differed remarkably between the Inner Mongolian and Tibetan grasslands. The means of both SIC and SOC in the Tibetan grasslands (0.51 % and 5.24 %, respectively) were higher ($P < 0.01$) than those in the Inner Mongolia grasslands (0.21 % and 1.61 %). However, SIC/STC had no significant difference between the Tibetan grasslands and the Inner Mongolian grasslands (0.18 and 0.11, $P = 0.54$). These patterns are also evident for the three soil depth increments (Table 3). Soil pH showed no significant difference between the two regions, while soils in the Tibetan grasslands generally accounted for higher soil total nitrogen and SM but lower SBD compared to the Inner Mongolian grasslands (Table 2).

On the whole, vegetation type had significant effects on SIC, SOC and SIC/STC ($P < 0.01$). Generally, SIC concentration was higher in alpine steppe soils than in meadow steppe and typical steppe soils. SOC was higher in alpine meadow soils compared to desert steppe soils. SIC/STC was larger as in alpine steppe soils as in the alpine meadow soils (Table 3).

3.2 Factors driving spatial variations of SIC and SOC

We developed regression tree models for SIC, SOC and SIC/STC (Fig. 3a–c). All three trees were significantly different from a random tree ($P < 0.05$). The trees for SIC, SOC and SIC/STC explained 69 %, 76 %, and 73 % of the variance, respectively. Tree models revealed that the spatial variations in SIC, SOC and SIC/STC were driven by very different factors (Table 4, Fig. 3).

For SIC (Fig. 3a), the tree model had an initial split on soil pH with a threshold of 7.0, implying that soil pH was the most important variable explaining the spatial variation of SIC. As pH increases above this threshold, MAT and PE represent two thermal factors that became common to affect SIC contents, indicating a negative relationship between

SIC and thermal condition. In areas with pH below 7.0, MAP and SM, both reflecting water conditions, were included in the model: SIC in the topsoil was typically lower on average in low-precipitation (<465 mm yr^{-1}) areas; in the areas with more precipitation, SIC was relatively high in soils with low moisture.

Vegetation type was the most important variable for SOC (Fig. 3b), explaining 49 % of the variations of SOC. Steppe soils had much lower SOC than meadow soils. In the steppe, only GSP influenced SOC, with a trend that SOC increased with GSP. In the meadow, AGB and PE were involved. Lower AGB (< 95.4 g m^{-2}) corresponded to lower SOC; but when AGB was above this threshold, SOC decreased with increasing of PE.

Soil pH was the initial split variable with a threshold of 7.0 for SIC/STC (Fig. 3c). When soil pH increases above this threshold, AGB and MAT entered the model, while SM and BGB entered the model when pH is below this threshold. These environmental factors impact the spatial pattern of SIC/STC through influencing SIC and/or SOC in the topsoil.

3.3 Empirical models for estimating the change of SIC

Regression trees showed that soil pH was the most important variable driving the pattern of both SIC and SIC/STC. Therefore we built linear regression models to predict SIC and SIC/STC from soil pH (Fig. 4a,b).

SIC and SIC/STC were both correlated with soil pH positively ($P < 0.01$). The model for SIC had a slope of 0.58 with an R^2 of 0.29 (Fig. 4a), while the model for SIC/STC had a slope of 0.75 with an R^2 of 0.43 (Fig. 4b). The analysis showed that it was adequate to estimate SIC in the topsoil with the models, considering the sample size and spatial pattern. The results of these empirical models showed that 1-unit decrease in soil pH would lead to 73 % and 82 % decrease in SIC and SIC/STC, respectively.

Furthermore, SIC/STC decreases faster than SIC when soil pH goes down, suggesting that SOC may also be affected by soil acidity. Thus we did linear regression analyses to test the relationship between SOC and soil pH. The result showed a significant

negative relationship between SOC and soil pH ($P < 0.05$), with a slope of -0.29 and an R^2 of 0.18 (Fig. 4c).

4 Discussion

4.1 Higher SIC and SOC in the Tibetan grasslands

Our results showed for both SIC and SOC significantly higher concentrations in the topsoil of the Tibetan grasslands than for those of the Inner Mongolian grasslands, at site level as well as for in each depth increment.

Across all the sites, SOC concentration is approximately 9 times as high as SIC concentration. The SIC content in our study, calculated with bulk density, had a mean value of 3.9 kg m^{-3} in 0–10 cm depth, which is lower than the average SIC content of 6 kg m^{-3} in the same depth increment in Chinese grasslands reported by Mi et al. (2008). Most likely, this is the case because the western Tibetan grasslands, which generally have higher SIC contents, were not included in our studies. Yang et al. (2010a) reported an average SIC density of 5.7 kg m^{-2} in the top 30 cm soil of the Tibetan grasslands, equal to 19.0 kg m^{-3} in 0–30 cm soil depth. Even considering different focused soil depth, their result is still much higher than the results of Mi et al. and the current study. The average SOC concentration in 0–20 cm soil depth in our study was 38.4 g kg^{-1}, close to the average value of 38.5 g kg^{-1} in Chinese grasslands reported by Xie et al. (2007).

Differences in soil formation and climatic conditions between Inner Mongolian grasslands and the Tibetan grassland may contribute to the above-described patterns. The fact that there was no significant difference between soil pH in Inner Mongolian and Tibetan grasslands suggests that higher SIC in the Tibetan grasslands may be due to basically two other reasons. Firstly, comparing with soils of the Inner Mongolian grasslands, soils of the Tibetan grasslands developed later thus parent material had stronger effects on soil characteristics, and carbonate migration is also lower (Xiong and Li, 1987). This would lead to higher LIC in the soil in the Tibetan grasslands. Secondly, due to higher elevation, CO_2 partial pressure is lower in the Tibetan grasslands (Körner, 2003). Lower temperature also induces lower soil respiration thus lower CO_2 partial pressure in the Tibetan grasslands soil (Kato et al., 2006). This influences the formation of pedogenic carbonate as presented in the following chemical equation:

$$Ca^{2+} + 2HCO_3^- \leftrightarrow CaCO_3 + H_2O + CO_2 \quad (1)$$

Consequently, lower CO_2 partial pressure will move the equilibrium towards more precipitation of carbonate (Nordt et al., 2000), benefitting the formation of PIC. Although this chemical reaction is well known, no studies considered CO_2 partial pressure as an important factor in determining the large-scale pattern of SIC. For the first time, our results showed that CO_2 partial pressure may play a key role in shaping topsoil SIC in extreme high altitudinal environments.

The pattern of SOC could be attributed to climatic differences between these two regions. Low temperature leads to slower decomposition rates (Wu et al., 2003; Xie et al., 2007; Kirschbaum, 1995), and high precipitation causes high vegetation productivity (Jobbagy and Jackson, 2000; Wynn et al., 2006; Callesen et al., 2003). In addition, high moisture, and especially temporal water saturation in the Tibetan grasslands due to the frozen ground, also leads to slower decomposition rates (Baumann et al., 2009). All these factors contribute to the accumulation of SOC in the topsoil of the Tibetan grasslands.

4.2 Different controls on the large scale patterns of SIC and SOC

According to the tree models, SIC and SOC in the topsoil were affected by different factors. The pattern of SIC in the topsoil could be well explained by climate, soil physical and chemical properties (including soil pH, PE, MAT, MAP and SM), while for SOC biotic and climatic factors were predominant (including vegetation type, AGB, PE and GSP).

In our research, soil pH is the most important factor for explaining the variation of SIC. Importantly, this strong relationship can be related to a decrease in HCO_3^- due to low pH as shown by the following equation:

$$CaCO_3 + 2H^+ \leftrightarrow Ca^{2+} + H^+ + HCO_3^- \leftrightarrow Ca^{2+} + CO_2 + H_2O \quad (2)$$

Acidification induces the equilibrium towards right, thus decreasing the formation of SIC (Lal and Kimble, 2000; Suarez, 2000). In our study, 42 % of the variation in SIC could be attributed to the differences of soil pH among the 81 sites. Within different pH range ($pH < 7.0$ or $pH > 7.0$), distinct environmental factors drive the finer pattern of SIC, indicating that SIC is influenced by different processes depending on soil pH. As pH increases above the threshold, thermal factors had stronger effects than other factors, showing negative relationships with SIC in the topsoil. This is consistent with results reported in other studies (Mi et al., 2008; Yang et al., 2010a). Temperature affects $CaCO_3$ equilibria through its effects on $CaCO_3$ solubility, evapotranspiration and leaching (Lal and Kimble, 2000), but it is not appropriate to attribute variation in SIC to temperature in our research. Temperature affects biological activity especially soil respiration positively (Davidson and Janssens, 2006; Raich and Schlesinger, 1992), which increases soil CO_2 partial pressure. Moreover, In our study region, temperature decreases with increasing elevation. Consequently, higher temperature means lower altitude and thus in turn higher CO_2 partial pressure. High CO_2 partial pressure would inhibit the formation of carbonates in the topsoil (Eq. 1), leading to negative relationships between SIC and thermal factors. In the areas with $pH < 7.0$, SIC tends to be in a form of dissolved bicarbonate and leaching process becomes more important, consequently water condition factors like MAP and SM entered the model. However, the result of our research showing that lower MAP is corresponding to lower SIC, is in contrast to some other studies, in which SIC shows a decreasing trend along precipitation gradients (Mi et al., 2008; Nordt et al., 2000; Wang et al., 2010; Yang et al., 2010a). Further analysis suggested that most of those sites with MAP < 465 mm were distributed in the Inner Mongolia, while those sites with MAP > 465 were all located on the Tibetan Plateau. Thus the positive SIC-precipitation relationship in this study may be caused by other confounding factors that were different between the two regions.

For SOC, steppe soils present much lower SOC contents than meadow soils. As the main resource of SOC, vegetation determines quantity, quality and distribution of SOC (Jobbagy and Jackson, 2000; Poeplau et al., 2011; Quideau et al., 2001). In Chinese grasslands, meadow has higher productivity than steppe (Ma et al., 2010; Ni, 2004), which means more organic matter input into meadow soils compared to steppe sites. Moreover, alpine meadows occur under cold and humid environment. Consequently, decomposition of SOC is limited by temperature and temporal water saturation (Davidson and Janssens, 2006; Jobbagy and Jackson, 2000; Baumann et al., 2009). These two interrelated processes (i.e. high input and low decomposition) both contribute to higher SOC in the topsoil of meadow ecosystem. Within different vegetation types, finer patterns of SOC were driven by different processes. In meadows, AGB showed a strong effect on SOC in the topsoil. Because AGB reflects vegetation productivity (LeBauer and Treseder, 2008; Ni, 2004), the latter influences SOC content in the topsoil of the meadow sites by controlling the input of organic matter. When AGB increases above the threshold, PE has a negative relationship with SOC. Although high temperature may stimulate plant production as temperature is an important limiting factor for vegetation growth in the alpine meadow (Piao et al., 2006), temperature-induced higher soil decomposition overrides the resulting increase in C inputs, thus leading to a decrease of SOC. On the other hand, in steppe ecosystems, water conditions are the primary constraint to vegetation productivity (Heisler-White et al., 2008; Jobbagy et al., 2002; Sala et al., 1988). High GSP could stimulate plant production and increase soil C accumulation, thus precipitation is the most important variable in controlling SOC patterns in steppe ecosystems.

4.3 Decrease in SIC but increase in STC under soil acidification

Our study provides insights in estimating future change of SIC under the scope of soil acidification scenarios. During the past several decades, parts of Europe and Eastern

North America suffered from significant acidification due to acid deposition (Bowman et al., 2008; Galloway et al., 2003). For example, in Europe soil pH decreased between 0.2 to 2 units over the past 17 to 110 yr (Stevens et al., 2009). China also has to face severe soil acidification problems due to increasing energy demand and excessive application of chemical fertilizer (Hicks et al., 2008; Larssen and Carmichael, 2000; Guo et al., 2010). Considering the important role of soil pH in controlling the pattern of SIC, soil acidification will have a great impact on topsoil's SIC stocks in the future.

Up to date, no studies have investigated the trend or extent of soil acidification in Chinese grasslands. However, soil pH records in croplands close to our study region provide circumstantial evidences to approximate the extent of soil acidification. From the 1980s to the 2000s, soil pH in croplands on the Loess Plateau and the Tibetan Plateau declined by 0.27 and 0.48 units, respectively (Zhou et al., 2009; Guo et al., 2010). Assuming the acidification rate in croplands is the same as that in grasslands and remains the same in the next 20 yr, according to our linear regression models, acidification will lead to 30 % and 53 % decrease in SIC in the Inner Mongolian and Tibetan grasslands, respectively.

Although acidification leads to decrease in SIC, carbon stock in the topsoil will not necessarily decline with soil acidification. It can be assumed, that due to accumulation of SOC, STC is likely to rise with increasing acidity, backed up by the negative relationship between soil pH and SOC as well as the even faster decline of SIC/STC compared to SIC with decreasing soil pH. This pattern can be explained by two reasons. Firstly, acidification inhibits soil microbial activities and thus the SOC decomposition rate (Francis, 1986). Secondly, N deposition, a major cause of acidification, will lead to a decrease in microbial biomass and oxidase activity (Dalmonech et al., 2010; Fisk and Fahey, 2001; Zak et al., 2008) and to an increase of SOC inputs through increasing vegetation productivity (Neff et al., 2002). Both of these will contribute to accumulation of SOC, thus leading to an increase in STC. Moreover, SIC only accounts for a very small proportion of STC on average. Therefore, soil acidification would not make a great impact on soil carbon.

4.4 Uncertainty of prediction

In the present study, the linear regression models with soil pH as independent variable were used to predict the future change of soil C under soil acidification scenario. Soil pH is the most important variable driving the large-scale pattern of SIC and SIC/STC. SOC also has a significant relationship with soil pH. These two facts suggest that our prediction could be helpful in understanding soil C change with acidification. However, some uncertainties remain in the prediction. The linear regression models were based on spatial relationships between soil pH and carbon content. Therefore, there may be some bias for predicting temporal soil carbon change using the models. Additionally, we used the acidification data in cropland near our study region since no other data is available to investigate the trend or extent of soil acidification in Chinese grasslands. This may also bring prediction errors.

Further studies are needed to validate the robustness of the relationships between soil carbon and soil pH at different spatial and temporal scales. In particular, long-term multi-factor experiments along environmental gradients might be useful to test these relationships. Moreover, the extents and trends of soil acidification in Chinese grasslands should also be investigated to give more accurate prediction of soil carbon change in the future.

5 Conclusions

Through analyzing the topsoil data of 81 sites at three different depth increments obtained from a regional survey across grasslands in Inner Mongolia and the Tibetan Plateau during 2006–2007, we found that both SIC and SOC in the topsoil in the Tibetan grasslands were significantly higher than those in the Inner Mongolian grasslands. Higher SIC in the Tibetan grasslands may be due to higher LIC derived from parent material and more PIC formation caused by lower CO_2 partial pressure, whereas higher SOC in the Tibetan grasslands is caused by higher litter input and

lower decomposition rates. At a large-scale, SIC and SOC were controlled by different environmental factors. SIC was mainly driven by chemical and physical processes, particularly by soil pH and other processes depending on soil pH. However, SOC was controlled by biotic processes such as vegetation type. Our results imply that given the acidification rate in Chinese grassland soils in the future is the same as that in Chinese cropland soils during the past two decades, SIC will decrease by 30 % and 53 % in the Inner Mongolian grasslands and the Tibetan grasslands respectively in the next 20 yr. However, the negative relationship between soil pH and SOC suggests that acidification will inhibit decomposition of SOC thus will not lead to a significant general loss of carbon from soils in these regions.

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Table 1. Description of the study regions. Mean annual temperature (MAT), mean annual precipitation (MAP), growing season temperature (GST), growing season precipitation (GSP), averaged topsoil pH, soil bulk density (SBD), soil moisture (SM) and soil total nitrogen (STN) of the sampling sites are shown.

	Overall	Inner Mongolian grasslands	Tibetan grasslands
No. of site	81	36	45
Longitude (° E)	90.80–120.12	111.83–120.12	90.80–101.48
Latitude (° N)	30.31–50.19	41.79–50.19	30.31–37.28
Altitude (m)	549–5105	549–1418	2925–5105
MAT (°C)	−5.8–4.1	−2.6–4.1	−5.8–2.6
GST (°C)	1.5–16.9	11.2–16.9	1.5–11
MAP ($mm\,yr^{-1}$)	148–604	148–436	218–604
GSP ($mm\,yr^{-1}$)	115–402	115–343	133–402
Soil pH	5.2–8.2	5.7–8.2	5.2–7.6
SBD ($g\,cm^{-3}$)	0.25–1.63	0.94–1.63	0.25–1.43
SM (V/V %)	0.73–68.11	2.04–16.10	0.73–68.11
STN (g/g %)	0.04–1.52	0.05–0.45	0.04–1.52

Table 2. Topsoil properties in the Inner Mongolian and Tibetan grasslands across sampling sites. Different letters indicate statistical significance at $P < 0.05$. SBD = soil bulk density; SM = soil moisture; STN = soil total nitrogen; CV = coefficient of variation.

		Inner Mongolian grasslands	Tibetan grasslands
Soil pH	*n*	36	45
	Mean	6.9**a**	6.8**a**
	CV	0.10	0.09
	Range	5.7–8.2	5.2–7.6
SBD ($g\,cm^{-3}$)	*n*	36	45
	Mean	1.3**b**	0.95**a**
	CV	0.13	0.33
	Range	0.94–1.63	0.25–1.43
SM (V/V %)	*n*	32	44
	Mean	7.29**a**	20.95**b**
	CV	0.55	0.85
	Range	2.04–16.10	0.73–68.11
STN (g/g %)	*n*	36	45
	Mean	0.17**a**	0.44**b**
	CV	0.54	0.81
	Range	0.05–0.45	0.04–1.52

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Table 3. Statistical description of soil inorganic carbon (SIC), soil organic carbon (SOC) content and SIC to soil total carbon ratio (SIC/STC) in each depth increment in different regions and vegetation types. Different letters indicate statistical significance at $P < 0.05$. CV = coefficient of variation.

		SIC (%)				SOC (%)				SIC/STC			
		0–5 cm	5–10 cm	10–20 cm	Site level	0–5 cm	5–10 cm	10–20 cm	Site level	0–5 cm	5–10 cm	10–20 cm	Site level
Overall	Mean	0.31	0.35	0.44	0.38	4.78	3.74	2.92	3.58	0.11	0.13	0.18	0.15
	CV	1.49	1.52	1.33	1.36	1.08	1.17	1.32	1.19	1.46	1.43	1.16	1.26
	Range	0.00–1.92	0.00–2.28	0.00–2.61	0.00–2.20	0.29–27.51	0.30–25.43	0.28–26.00	0.32–26.34	0.00–0.66	0.00–0.66	0.00–0.74	0.00–0.65
	n	87	87	87	81	87	87	87	81	87	87	87	81
Region													
Inner Mongolian grasslands	Mean	0.14**a**	0.18**a**	0.25**a**	0.22**a**	2.23**a**	1.58**a**	1.26**a**	1.61**a**	0.07**a**	0.09**a**	0.14**a**	0.11**a**
	CV	2.09	2.30	1.81	1.89	0.72	0.66	0.65	0.66	1.44	1.66	1.31	1.39
	Range	0.01–1.59	0.00–2.12	0.01–2.37	0.01–2.10	0.29–7.54	0.30–4.42	0.28–3.60	0.32–4.55	0.00–0.39	0.00–0.53	0.00–0.65	0.00–0.54
	n	39	39	39	36	39	39	39	36	39	39	39	36
Tibetan grasslands	Mean	0.45**b**	0.49**b**	0.58**b**	0.51**b**	6.84**b**	5.50**b**	4.27**b**	5.15**b**	0.15**a**	0.17**a**	0.21**a**	0.18**a**
	CV	1.18	1.19	1.07	1.10	0.89	0.95	1.11	0.99	1.30	1.27	1.06	1.16
	Range	0.00–1.92	0.00–2.28	0.00–2.61	0.00–2.20	0.41–27.51	0.35–25.43	0.45–26.00	0.45–26.34	0.00–0.66	0.00–0.66	0.00–0.74	0.00–0.65
	n	48	48	48	45	48	48	48	45	48	48	48	45
Vegetation type													
Meadow steppe	Mean	0.06**ab**	0.04**ab**	0.16**ab**	0.11**a**	3.21**b**	2.46**b**	2.04**b**	2.43**b**	0.03**a**	0.02**a**	0.09**ab**	0.06**ab**
	CV	0.88	0.47	1.71	1.41	0.54	0.57	0.53	0.53	1.21	0.68	1.49	1.46
	Range	0.02–0.19	0.02–0.07	0.01–0.78	0.01–0.44	0.88–5.29	0.73–4.17	0.59–3.60	0.70–4.15	0.00–0.11	0.00–0.05	0.00–0.37	0.00–0.25
	n	7	7	7	7	7	7	7	7	7	7	7	7
Typical steppe	Mean	0.16**a**	0.21**a**	0.28**a**	0.24**a**	2.42**b**	1.64**ab**	1.25**ab**	1.65**ab**	0.05**a**	0.08**a**	0.12**ab**	0.09**ab**
	CV	2.23	2.36	1.93	2.02	0.64	0.52	0.51	0.55	1.81	1.79	1.46	1.52
	Range	0.01–1.59	0.00–2.12	0.01–2.37	0.01–2.10	0.69–7.54	0.57–4.418	0.48–3.51	0.59–4.55	0.00–0.39	0.00–0.50	0.01–0.63	0.01–0.51
	n	24	24	24	22	24	24	24	22	24	24	24	22
Desert steppe	Mean	0.23**ab**	0.31**abc**	0.44**ab**	0.39**ab**	1.07**a**	0.98**a**	0.80**a**	0.98**a**	0.17**b**	0.20**ab**	0.26**ab**	0.24**bc**
	CV	1.08	1.19	1.11	1.00	0.89	0.82	0.75	0.73	0.81	0.95	0.75	0.79
	Range	0.01–0.76	0.01–0.92	0.02–1.25	0.01–0.94	0.29–3.40	0.30–2.96	0.28–2.07	0.32–2.62	0.02–0.45	0.02–0.53	0.02–0.65	0.02–0.54
	n	10	10	10	9	10	10	10	9	10	10	10	9
Alpine steppe	Mean	0.71**c**	0.76**c**	0.83**b**	0.78**b**	2.42**b**	2.07**ab**	1.80**b**	1.97**ab**	0.29**b**	0.30**b**	0.34**b**	0.32**c**
	CV	0.82	0.90	0.89	0.88	0.72	0.63	0.61	0.65	0.75	0.77	0.68	0.69
	Range	0.00–1.92	0.04–2.28	0.01–2.61	0.02–2.20	0.41–6.24	0.35–4.33	0.45–4.00	0.45–4.19	0.00–0.66	0.01–0.66	0.01–0.74	0.01–0.65
	n	18	18	18	17	18	18	18	17	18	18	18	17
Alpine meadow	Mean	0.27**b**	0.31**bc**	0.39**ab**	0.30**ab**	10.02**c**	7.94**c**	6.03**c**	7.47**c**	0.05**a**	0.08**a**	0.11**a**	0.07**a**
	CV	1.63	1.51	1.27	1.25	0.61	0.71	0.92	0.76	2.01	2.01	1.53	1.66
	Range	0.00–1.698	0.00–1.75	0.00–1.71	0.00–1.68	1.07–27.51	0.82–25.43	0.69–26.00	0.81–26.34	0.00–0.40	0.00–0.63	0.00–0.67	0.00–0.55
	n	28	28	28	26	28	28	28	26	28	28	28	26

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Table 4. Variables included and selected in the regression tree analysis. Alt = altitude; MAT = mean annual temperature; GST = growing season temperature; MAP = mean annual precipitation; GSP = growing season precipitation; PE = potential evapotranspiration; AE = actual evapotranspiration; AP/PE = the ratio of MAP to PE; VT = vegetation type; AGB = aboveground biomass; BGB = belowground biomass; SBD = soil bulk density; SM = soil moisture.

				Variables used to build the tree models			Variables selected finally		
Variable	*n*	Mean	SD	SIC	SOC	SIC/STC	SIC	SOC	SIC/STC
Alt	80	2692	1692.4	Yes		Yes			
MAT	79	−1.1	2.42	Yes	Yes	Yes	Yes		Yes
GST	75	8.8	4.63	Yes	Yes	Yes			
MAP	79	384.0	113.66	Yes	Yes	Yes	Yes		
GSP	75	270.5	63.01	Yes	Yes	Yes		Yes	
PE	75	432.0	92.30	Yes	Yes	Yes	Yes	Yes	
AE	75	321.7	59.81	Yes	Yes	Yes			
AP/PE	74	1.0	0.43	Yes		Yes			
VT	81	–	–	Yes	Yes	Yes		Yes	
AGB	77	96.4	59.41	Yes	Yes	Yes		Yes	Yes
BGB	73	113.3	1515.15	Yes	Yes	Yes			
pH	81	6.88	0.563	Yes		Yes	Yes		Yes
SBD	81	1.11	0.315	Yes					
SM	76	15.2	15.27	Yes		Yes	Yes		Yes

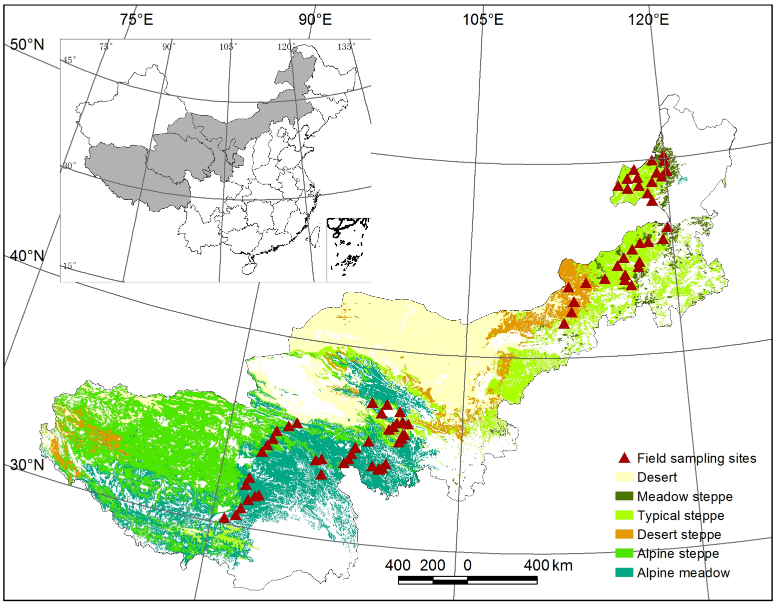

Fig. 1. Vegetation map of Chinese grasslands and location of sampling sites (1 : 1 000 000) (Chinese Academy of Sciences, 2001).

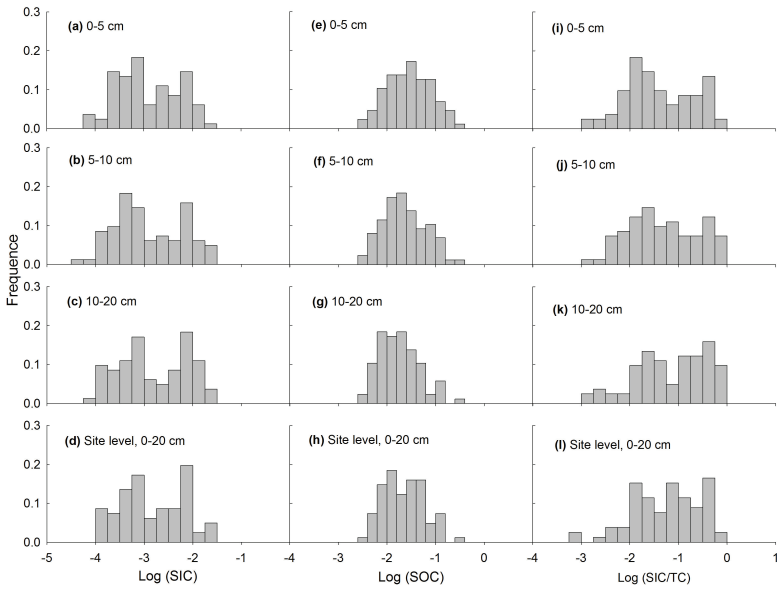

Fig. 2. Frequency distributions of SIC **(a–d)**, SOC **(e–h)** and SIC/STC **(i–l)** in Chinese grasslands across all sampling sites at soil depths of 0–5 cm **(a, e, i)**, 5–10 cm **(b, f, j)**, 10–20 cm **(c, g, k)** and at site level **(d, h, l)**. All distributions have no significant differences compared with normal distribution at $P < 0.05$.

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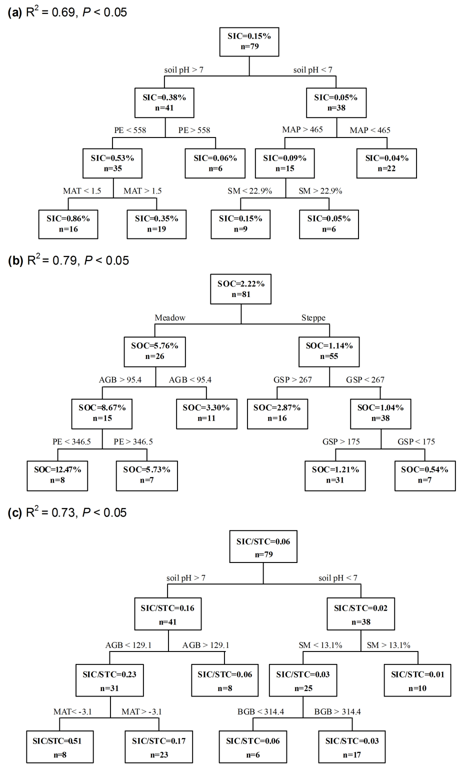

Fig. 3. Classification and regression trees of SIC **(a)**, SOC **(b)**, SIC/STC **(c)**. Branches are labelled with criteria used to segregate data. Values in terminal nodes represent mean soil SIC **(a)**, SOC **(b)** and SIC/STC (0–20 cm) of sites grouped within the cluster.

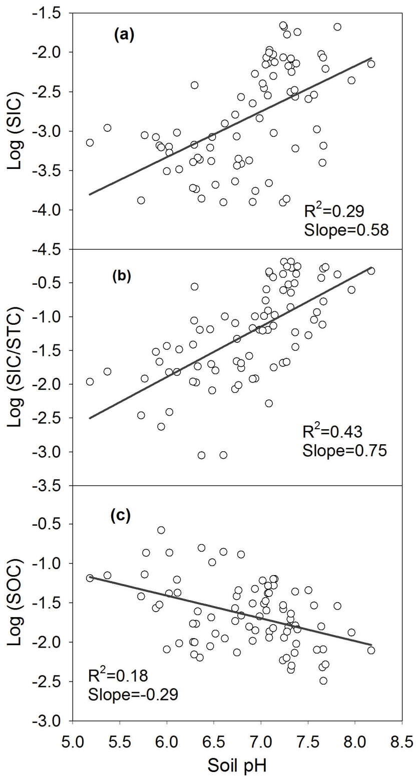

Fig. 4. Linear relationships of SIC **(a)**, SIC/STC **(b)** and SOC **(c)** with soil pH value in the topsoil. All regression lines were significant ($P < 0.05$).