Effect of climate change on soil organic carbon in Inner Mongolia

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ABSTRACT: This study simulated the response of soil organic carbon (SOC) to future climate change in Inner Mongolia. The Lund–Potsdam–Jena model, which is a dynamic vegetation model, was improved in accordance with the ecosystem features in China. In this study, this model was driven by the A1B, A2, and B2 climate scenarios generated by a regional climate model. To assess the spatial and temporal variations in SOC, four 30-year periods were defined, namely baseline term (1961–1990), near term (2011–2040), middle term (2041–2070), and long term (2071–2100). Results suggest that, under the climate scenarios, soil organic carbon density (SOCD) significantly decreases from 1991 to 2100. SOCD may slightly decrease before 2025 but rapidly decreases after 2025, particularly for the A1B scenario, which shows the largest decrement in SOCD, followed by the B2 and A2 scenarios. Large and small decrements of SOCD may occur in Western and Eastern Inner Mongolia, respectively. The effects of climate change on SOCD are generally insignificant in the near term. In the middle term, the areas of SOCD decrement would expand, and the negative effects of climate change would dominate. In the long term, the negative effect of climate change would be further enhanced, and the SOCD in most areas of Inner Mongolia would decrease.

KEY WORDS soil organic carbon; dynamic response; climate change; Inner Mongolia

Received 20 August 2013; Revised 11 February 2014; Accepted 13 February 2014

1. Introduction

The interaction between climate warming and the global carbon cycle is an important aspect of global environmental changes (IPCC, 2007). The ecosystem carbon stock has an irreplaceable function in mitigating climate change as a key component in the global carbon cycle. The soil organic carbon (SOC) stock is the largest carbon stock in the terrestrial ecosystem at thrice the global atmospheric carbon stock and 1.5–2 times the vegetation carbon stock, which accounts for two thirds of the total terrestrial ecosystem carbon stock (Post *et al.*, 1982; Schlesinger, 1990). Therefore, changes in SOC storage significantly influence climate change, and a slight change in SOC stocks can have a considerable effect on atmospheric CO₂ concentration, which contributes to climate warming (Davidson and Janssens, 2006). Climate warming will accelerate SOC decomposition and enhance vegetation productivity, which increases litter in the soil, thus offsetting a portion of SOC losses. However, minimal information has been revealed about the trends and future projected changes in SOC in China (Trumbore and Czimczik, 2008). Thus, the accurate estimation of SOC dynamics is important to evaluate the influence of climate change on a terrestrial ecosystem.

Grassland ecosystems are the most widely distributed ecosystems in the world and cover ~40.5% of the global land surface. Grassland ecosystems serve an important function in the global carbon cycle. A considerable part of CO₂ emissions are sequestered by grassland ecosystems, which account for 23% of the global terrestrial ecosystem carbon storage (WBGU, 1998). Approximately 89% of the carbon sequestered by grasslands is stored in soil (Ajtay *et al.*, 1979). Studies on SOC have been extensively undertaken in different regions (Raich and Schlesinger, 1992; Ojima *et al.*, 1993; Zhang YQ *et al.*, 2006; Zhang Y *et al.*, 2006; Xie *et al.*, 2007). The changes in and responses of SOC to climate change have also received considerable attention. The temporal and spatial changes in SOC at the regional scale are difficult to estimate by field surveys and observation (Smith, 2004). Model simulation is important in examining SOC responses to climate change because this approach can simulate the interaction and feedback among different ecosystem processes. This approach is also important for understanding the nutrient cycles, energy flows, and carbon exchanges in an ecosystem (Cramer *et al.*, 2001; Bellamy *et al.*, 2005).

Parton *et al.* (1995) have estimated global grassland soil carbon when air temperature increases by 2–5 °C and predicted that SOC in grasslands will lose 3–4 Pg C in 50 years because of the increased SOC decomposition rates attributed to global warming. Riedo *et al.* (2000) indicated that carbon is typically lost from grazed grassland with a 4 °C increase in temperature. Xiao *et al.* (1996)

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examined the effect of global climate change and suggested that climate change would result in a considerable decrease in soil organic matter (SOM) of *Aneurolepidium chinense* and *Stipa grandis* steppe. Wan *et al.* (2011) modelled the effects of climate change on SOC stock in upland soils in China. They reported that SOC would decrease in most areas of China in the 21st century, especially in Northern China. In these studies, biogeochemical models, which simulate fluxes of carbon, water, and nitrogen coupled in an ecosystem by assuming a constant distribution of vegetation, are usually selected as the main approach to examine SOC responses to climate change. However, the vegetation dynamic is important when assessing changes in carbon storage because rapid climate change can result in significant changes in vegetation distribution. Therefore, projecting changes in SOC under rapid climate change requires a more comprehensive model that includes not only biogeochemical processes but also vegetation dynamics.

Inner Mongolia, which is located in Northern China, is the hinterland of the Eurasian continent. Inner Mongolia is strongly influenced by the Asian monsoon climate with increasing aridity from the northeast to the southwest. Three main vegetation types are distributed along the gradient in this area. These types are forests, grasslands, and deserts, which provide opportunities for the examination of SOC changes in various types of biomes. Most lands in Inner Mongolia are covered by natural grasslands that span ~ 1.18 million km², accounting for one third of China's grassland area (Gao and Fan, 1997). The grassland ecosystem in Inner Mongolia has high plant biodiversity and is sensitive to climate changes because of specific geographical positions and semi-arid and arid climate conditions (Niu, 2001). In recent decades, climate change and anthropogenic activities have highlighted the vulnerability and sensitivity of the ecosystem in Inner Mongolia (Xu *et al.*, 2009; Zuo *et al.*, 2009; Ma *et al.*, 2010). Ecological research in Inner Mongolia has drawn significant attention, particularly after the implementation of the Grain to Green Program and Grazing Withdrawal Program by the Chinese central government in 2000.

Vegetation dynamics, biogeography, and biogeochemical processes interact with one another and are inseparable in ecosystem responses to climate change. However, the integrated consideration of the two aspects is not comprehensive in previous studies. A dynamic vegetation model, which combines vegetation dynamics, biogeography, and biogeochemical processes in a modular framework, can simulate the responses of an ecosystem to climate change in different temporal and spatial scales. By using the Lund–Potsdam–Jena (LPJ) dynamic vegetation model (Sitch *et al.*, 2003), which was improved based on the features of the Chinese ecosystem, this study simulated SOC in Inner Mongolia based on regional climate scenarios (A2, B2, and A1B). The temporal and spatial changes in SOC in different regions were examined based on eco-regions. This study aimed to elucidate the effects of climate change on SOC at the regional scale and support

local government units in their efforts to mitigate climate change.

2. Data and methods

2.1. Climate data

Climate scenario data were provided by the Institute of Environment and Sustainable Development in Agriculture of the Chinese Academy of Agricultural Sciences. This study group generated the climate scenarios in the 21st century across China with a spatial resolution of 50 km $\times$ 50 km by using the Providing Regional Climate for Impacts Studies (PRECIS) system (Jones *et al.*, 2004) based on the Special Report on Emission Scenarios (SRES) (Nakicenovic *et al.*, 2000). A general circulation model, i.e. UKMO_HadCM3, was used to obtain the boundary conditions for PRECIS. UKMO_HadCM3 is considered more effective than other general circulation models in simulating the spatial and temporal characteristics of climate change in China (Miao *et al.*, 2012). PRECIS has been validated to simulate surface climate variations efficiently (Xu and Jones, 2004; Xu *et al.*, 2006; Zhang YQ *et al.*, 2006; Zhang Y *et al.*, 2006) and has been extensively applied to various impact assessments of climate change in China (Xiong *et al.*, 2009; Zhao *et al.*, 2011a, 2011b). This study selected climate data based on the A1B ('global-economic'), A2 ('regional-economic'), and B2 ('regional-environmental') scenarios from SRES to evaluate the effects of climate change on the soil organic carbon density (SOC_D) of Inner Mongolia. The projected climate data used in this study included the maximum, minimum, and average temperatures, as well as precipitation, mean relative humidity, mean wind speed, and cloudiness.

On the basis of the projection of PRECIS, the annual mean temperature of Inner Mongolia will increase by 5.0 °C under the A1B scenario and 3.5 °C under the B2 scenario at the end of this century compared with the baseline term (Figure 1) (1961–1990). The annual mean temperature will increase from west to east under the A2 and B2 scenarios. In the A1B scenario, the increase in temperature in Eastern Inner Mongolia will be the maximum temperature, followed by the temperature in Western Inner Mongolia. The minimum temperature will be located in Central Inner Mongolia (Figure 2). Annual precipitation will also increase with large inter-annual variations, particularly in the B2 scenario. At the end of this century, precipitation is projected to increase by 24% in the A2 scenario and 14% in the B2 scenario. The increase in precipitation will also show large spatial heterogeneity. The overall trend shows that precipitation will be reduced from Western Inner Mongolia to Eastern Inner Mongolia (Figure 3).

To study the spatial and temporal variations in SOC, four 30-year periods were divided as follows: baseline term (1961–1990), near term (2011–2040), middle term (2041–2070), and long term (2071–2100).

Figure 1. Inter-annual variations of temperature (a) and precipitation (b) in Inner Mongolia relative to the averages of the baseline term (1961–1990).

2.2. Soil data

This research adopted soil texture data from Shangguan *et al.* (2012), who produced China Soil Texture data sets based on a 1 : 1 000 000 soil map of China and relevant data from the second national soil survey. This data set contains information on the geographical distribution of different soil texture types. By coordinate transformation and resampling with ArcGIS 9.3 (ESRI, Redlands, CA, USA), the data were transformed into a raster data set in the ArcInfo Grid format with a spatial resolution of 50 km × 50 km grid to meet the input requirements of the LPJ model.

2.3. Methods

The LPJ model (Sitch *et al.*, 2003; Gerten *et al.*, 2004) is a moderately complex dynamic vegetation model that was developed based on the early equilibrium model (i.e. BIOME3). The LPJ model adopted several features of the BIOME series model. For example, the establishment and death of natural vegetation are mainly controlled by vegetation environmental limiting factors, and the gross primary productivity of each plant functional type (PFT) is calculated based on carbon–water coupling. LPJ combines the process-based representations of terrestrial vegetation dynamics and the land–atmosphere carbon–water cycle in a modular framework to simulate the cycling of carbon and nutrients in plants, soils, and the atmosphere. In the LPJ model, each PFT has values that are above and below the ground litter pools. A part of the litter is decomposed and transformed into CO₂ and then released into the atmosphere. The remaining

litter is divided into intermediate and slow SOM pools. The litter, medium-speed, and slow-speed pools have a standard decomposition rate that corresponds to the turnaround times in the three pools. Decomposition rate (k) is the function of soil temperature and moisture. Temperature dependence [$g(T)$] follows the modified Arrhenius relationship (Lloyd and Taylor, 1994). The soil moisture relationship [$f(w)$] was obtained from Foley (1995). The decomposition rate, temperature dependence, and soil moisture relationship are provided as follows:

$$k = \frac{(1/\tau_{10}) g(T) f(w)}{12}$$

$$g(T) = \exp \left[308.56 \cdot \left(\frac{1}{56.02} - \frac{1}{(T + 46.02)} \right) \right]$$

$$f(w) = 0.25 + 0.75w_1$$

$$c = c_0 \cdot e^{-kt}$$

where τ_{10} is the decomposition rate for SOM at 10 °C; c is the pool size at any time t , with c_0 representing its initial size; T is the soil temperature; and w_1 is the average moisture status in the upper soil layer.

In our previous study, LPJ was carefully modified by adding shrub and cold grass PFTs, which were parameterized based on various inventory and observational data in accordance with the characteristics of Chinese ecosystems (Zhao *et al.*, 2011a). The radiation-calibrated Penman–Monteith model was also adopted to estimate

Figure 2. Spatial changes of annual mean temperature in near term (2011–2040), middle term (2041–2070), and long term (2071–2100) compared with the baseline term projected by the PRECIS under A1B, A2, and B2 scenarios.

potential evapotranspiration in the LPJ model (Yin *et al.*, 2013). The simulated results by the improved LPJ for China (LPJ-CN) were validated by comparisons with data sets obtained from the observed sites (Zhao *et al.*, 2011a). The simulated net primary productivity (NPP) results from the LPJ-CN match the results of the observed data ($R^2 = 0.64$, $p < 0.01$), which are better than the original LPJ data ($R^2 = 0.10$) used by Ni (2003).

The LPJ-CN was implemented for a 1000-year spin-up period driven by the average climate data of 1961–1990 to reach equilibrium with respect to vegetation covers and carbon pools. After reaching equilibrium, LPJ was driven by transient climate data from 1961 to 2100 to execute a dynamic simulation.

2.4. Eco-geographical regions

The climate condition in Inner Mongolia is complex and varies from cold temperatures in the north to warm temperatures in the south and from humid in the east to dry in the west. The Inner Mongolian ecosystem is diverse with forests in the eastern region, grasslands in the central region, and deserts in the western region. On the basis of the eco-geographical regionalization

scheme from Zheng (2008), Inner Mongolia was divided into 12 eco-geographical regions with similar terrains, climates, and vegetation types (Figure 4): cold temperate humid, deciduous–coniferous forest region of the Northern Greater Hinggan Mountain (IA1); medium temperate humid, broadleaved and coniferous-mixed forest region of the Piedmont Platform of the Eastern Songliao Plain (IIA3); medium temperate sub-humid, forest–steppe region of the Central Songliao Plain (IIB1); medium temperate sub-humid, forest–steppe region of the Central Greater Hinggan Mountain (IIB2); medium temperate sub-humid, forest–steppe region of the hill land of the Northwestern Greater Hinggan Piedmont (IIB3); medium temperate semi-arid, plain–steppe region of the Western Liaohe River (IIC1); medium temperate semi-arid, steppe region of Southern Greater Hinggan Mountain (IIC2); medium temperate semi-arid, steppe region of Eastern Inner Mongolia highland (IIC3); medium temperate semi-arid, steppe region of the Hulun Buir Plain (IIC4); medium temperate arid, desert–steppe region of Ordos and Western Inner Mongolia highland (IID1); medium temperate arid, desert regions of Alax and Hexi corridor (IID2); and warm temperate humid,

Figure 3. Spatial changes of annual mean precipitation in near term (2011–2040), middle term (2041–2070), and long term (2071–2100) compared with the baseline term projected by the PRECIS under A1B, A2, and B2 scenarios.

deciduous–broadleaved forest region of mountains in North China (IIIB3). The eco-geographical regions were used as a framework for analysing the regional differences of SOCD spatiotemporal changes.

3. Results

3.1. Evaluation of simulated results

The simulated results were evaluated through comparisons with the observed data. The observed SOCD data used in the accuracy evaluation included two parts: one part was from the North China grassland SOCD data obtained by five consecutive annual soil surveys from 2001 to 2005 on the grasslands of China (Yang *et al.*, 2010a, 2010b), and the other part was from the National Soil Inventory of China (IMSUI, 1994) undertaken during the 1980s. These inventory data comprise geographical location information, land cover type, and physical and chemical properties for different soil types. Soil inventory data are often used to study SOC spatial distributions, estimate SOC stock size, and evaluate the simulated results of SOC (Xiao *et al.*, 1996; Wang and Zhao, 1998; Zhang YQ *et al.*, 2006; Zhang Y *et al.*, 2006; Ji *et al.*, 2009). The SOCD simulated by LPJ-CN is closely correlated

with the observed data with a determination coefficient (R^2) of 0.716 (Figure 7). Figure 4 illustrates the spatial distribution of the SOCD simulated by LPJ-CN, which is consistent with the spatial distribution of the SOCD generated by Wang *et al.* (2003), Chen *et al.* (2003), and Xie *et al.* (2004) based on the national soil inventory data. These comparisons indicate that the improved LPJ model can accurately simulate the SOCD in Inner Mongolia and is reliable for regional SOCD assessments.

In the baseline period (1961–1990), the SOCD distribution in Inner Mongolia gradually decreased from 23 kg C m^{-2} in the northeast to 0.3 kg C m^{-2} in the southwest (Figure 5). The spatial characteristics of SOCD are relevant to vegetation distribution. The vegetation types of Inner Mongolia vary from deciduous–coniferous forests in the northeast to meadow steppes, typical steppes, desert steppes, and deserts in the southwest. This variation decreases the NPP of vegetation, thus resulting in the spatial difference in the litter input in the soil. The litter input differences, along with the regional temperature and precipitation, form the SOCD distribution pattern in Inner Mongolia.

Figure 4. Eco-geographical regions in Inner Mongolia. IA1 refers to cold temperate humid, deciduous–coniferous forest region of the Northern Greater Hinggan Mountain; IIA3 refers to medium temperate humid, broadleaved and coniferous-mixed forest region of the Piedmont Platform of the Eastern Songliao Plain; IIB1 refers to medium temperate sub-humid, forest–steppe region of the Central Songliao Plain; IIB2 refers to medium temperate sub-humid, forest–steppe region of the Central Greater Hinggan Mountain; IIB3 refers to medium temperate sub-humid, forest–steppe region of the Hill land of the Northwestern Greater Hinggan Piedmont; IIC1 refers to medium temperate semi-arid, plain–steppe region of the Western Liaohe River; IIC2 refers to medium temperate semi-arid, steppe region of Southern Greater Hinggan Mountain; IIC3 refers to medium temperate semi-arid, steppe region of Eastern Inner Mongolia highland; IIC4 refers to medium temperate semi-arid, steppe region of the Hulun Buir Plain; IID1 refers to medium temperate arid, desert–steppe region of Ordos and Western Inner Mongolia highland; IID2 refers to medium temperate arid, desert regions of Alax and Hexi corridor; IIB3 refers to warm temperate humid, deciduous–broadleaved forest region of mountains in North China.

3.2. SOCD changes in the regional scale

The simulated average SOCD in Inner Mongolia in the baseline term is 7.89 kg C m^{-2} , which is higher than the estimated average SOCD of 4.11 kg C m^{-2} based on the soil survey data by Yang *et al.* (2010a, 2010b). This difference is attributed to the forests distributed in North-eastern Inner Mongolia with high SOCD ($<20 \text{ kg C m}^{-2}$), thus increasing the average SOCD at the regional scale. The simulated SOCD in Inner Mongolia is lower than the estimated SOCD based on the national soil inventory data ($10.21 \text{ kg C m}^{-2}$) by Chen *et al.* (2003). This result is mainly attributed to the quantity of the sampling points, which is less in the west than in any other region in the national soil inventory, particularly in desert ecosystems with low SOCD. Thus, a high SOCD is estimated in Inner Mongolia. The simulated average SOCD of Inner Mongolia in the period 1961–1990 generally shows a decreasing trend (Figure 6) at $\sim 0.004 \text{ kg C m}^{-2} \text{ year}^{-1}$. The difference between the maximum and minimum values over 30 years is $<0.1 \text{ kg C m}^{-2}$, which is only 1.4% of the average value for the past 30 years. Given that the slight changes in SOCD are difficult to investigate by site-level comparisons, some studies usually

Figure 5. Simulated spatial distribution of the average SOCD from 1961 to 1990 in Inner Mongolia.

conclude that the SOCD in recent decades has remained constant.

In future climate scenarios, the average SOCD in Inner Mongolia may decrease (Figure 6). A considerable decrease may occur in the A1B scenario with a decreasing rate of $0.019 \text{ kg C m}^{-2} \text{ year}^{-1}$ ($R^2 = 0.94$), followed by the B2 scenario ($0.014 \text{ kg C m}^{-2} \text{ year}^{-1}$, $R^2 = 0.95$). The decrease in the average SOCD in the A2 scenario is the smallest and is $\sim 0.011 \text{ kg C m}^{-2} \text{ year}^{-1}$ ($R^2 = 0.85$). In climate scenarios, the SOCD changes can be divided into two periods, i.e. before and after 2025. The change in SOCD amplitudes before 2025 is small in the A1B and A2 scenarios. The SOCD may also exhibit a significant reduction after 2025. Compared with the baseline term (Figure 7), the decrease in SOCD in the A1B, A2, and B2 scenarios can reach 13, 7, and 9%, respectively. The conjunctive decrease in SOCD indicates that the carbon sequestration capacity of the soil in the Inner Mongolian ecosystem may decrease with climate change.

3.3. Spatial variation of SOCD

In the near term, the changes in SOCD in Inner Mongolia will be insignificant. Under the A2 and B2 scenarios, SOCD may increase in the east but decrease in the west (Figure 8). In the A1B scenario, SOCD shows different trends, i.e. a decreasing trend in the east and an increasing trend in the west. SOCD increments are mainly concentrated in the IIC3 and IID1 regions with amplitudes of 1.48 and 10.18%, respectively, compared with the baseline term (Table 1). The increase in SOCD can be associated with the spatial difference of precipitation variation in the near term. In the A1B scenario, precipitation may evidently increase in Western Inner Mongolia, wherein water is the main limitation to plant growth because the ecosystems in this region are dominated by steppes and desert steppes. An increase in precipitation can stimulate plants to produce more litter input to soil. Areas dominated by humid forests and sub-humid forests–steppes include the

Figure 6. Inter-annual variation of SOCD in Inner Mongolia from 1961 to 1990 in baseline term, and from 1991 to 2100 under A1B, A2, and B2 scenarios.

Figure 7. Comparison between the simulated and measured SOCD in Inner Mongolia.

IA1, IIA3, IIB1, IIB2, and IIB3 regions, wherein SOCD is 4% lower than the baseline term. However, the SOCD in semi-arid steppes and arid deserts is expected to be 5% higher than other regions, particularly in the IID1 region.

In the middle term, the SOCD in Inner Mongolia will decrease compared with the near term. In the A1B scenario, the SOCD in the Western Inner Mongolia may be enhanced, but areas with SOCD increments may shrink. In the A2 and B2 scenarios, SOCD may further decrease in the west and SOCD exceeds 6 and 14% in the IID1 and IID2 regions, respectively, compared with the

baseline term (Table 1). The IIC3 region has a concentrated distribution area of typical steppes, where SOCD exhibits varied trends under the three climate scenarios. In the A1B scenario, SOCD may increase and decrease in the west and east of IIC3, respectively. However, opposite variations are observed in the A2 scenario, wherein SOCD increases in the east and decreases in the west. In the B2 scenario, the overall SOCD tends to decrease (Figure 9). In the IA1, IIA3, IIB1, IIB2, and IIB3 regions, the average SOCD decreases in the A1B and B2 scenarios compared with the average SOCD in the A2 scenario. This difference is caused by the spatial heterogeneity to climate change. Compared with the near term, the area with reduced SOCD may increase, particularly in the A1B and B2 scenarios. The SOCD gradually decreases with climate change.

In the long term, the SOCD in Inner Mongolia will decrease and the spatial heterogeneity of SOCD changes will weaken (Figure 10). In the A2 scenario, the SOCD may increase in the middle area dominated by steppes. SOCD further decreases in the IID1 and IID2 regions, wherein the average decrement of SOCD exceeds 19% in the B2 scenario and 14% in the A1B scenario (Table 1). The decrease in SOCD in the IA1, IIA3, IIB1, IIB2, and IIB3 regions is twice the increase in the middle term, especially in the A2 scenario. The SOCD in the IIB3 region may decrease in the A1B scenario and increase in the A2 and B2 scenarios. Compared with the middle term, more eco-geographical regions may experience SOCD

Figure 8. SOCD spatial variation in near term (2011–2040), compared with baseline term, under A1B, A2, and B2 scenarios.

Table 1. Average SOCD of each eco-geographical region in the baseline term (1961–1990) compared with the SOCD in the near term (2011–2040), middle term (2041–2070), and long term (2070–2100) under the A1B, A2, and B2 scenarios.

Eco-geographical region	Baseline year (kg C m ⁻²)	Relative change (%)								
		Near term			Middle term			Long term		
		A1B	A2	B2	A1B	A2	B2	A1B	A2	B2
I A1	23.01	−1.81	−0.62	−0.99	−5.52	−3.35	−3.22	−11.20	−9.33	−7.21
II A3	18.72	−2.46	0.65	−0.63	−10.07	−2.28	−4.62	−22.52	−12.20	−13.14
II B1	11.16	−3.55	3.13	0.37	−14.02	−1.05	−6.99	−27.34	−10.25	−19.58
II B2	20.61	−2.74	−0.31	−3.14	−8.48	−3.20	−8.24	−17.55	−10.33	−15.70
II B3	19.82	−3.64	−0.33	−3.15	−10.33	−4.88	−7.75	−19.73	−13.10	−14.69
II C1	8.36	−0.34	7.79	2.09	−9.75	6.03	−4.98	−21.25	−1.08	−14.26
II C2	10.59	−3.29	6.98	1.45	−12.23	5.17	−4.08	−22.79	−1.40	−10.53
II C3	7.07	1.48	0.30	−6.29	−3.52	2.30	−8.67	−14.80	2.23	−14.68
II C4	9.22	−8.87	2.70	−6.64	−22.63	−6.88	−15.79	−36.04	−19.79	−27.26
II D1	1.04	10.18	−9.68	−5.65	5.71	−6.12	−7.81	−8.40	2.02	−13.76
II D2	0.34	−2.82	−13.89	−8.29	−12.12	−14.73	−14.76	−19.62	−6.98	−25.14
III B3	11.46	−0.30	5.97	13.01	−6.28	1.85	12.84	−15.07	−4.99	8.01

reductions in the long term, and the increase in SOCD may be concentrated in Central and Western Inner Mongolia with low SOCD values and large percentage changes.

4. Discussion

The simulated spatial distribution of SOCD in Inner Mongolia decreases from the northeast to the southwest. This trend is consistent with the spatial distribution of precipitation and opposite to the spatial changes in temperature. This change supports the views of Post *et al.* (1990), which indicate that SOCD increases with increasing precipitation and decreasing temperature. SOC is a balanced result of organic matter input *versus* its loss. Climate change affects the sequestration and decomposition of SOC through altering the litter input and decomposition rates of SOC. Climate in Inner Mongolia projected changes characterized by rising temperature and slightly increasing precipitation in the future. Considerable research demonstrated that climate warming can increase soil temperature, thus accelerating microbial decomposition (Friedlingstein *et al.*, 2001; Lin *et al.*, 2011), and can stimulate plant production by prolonging the growing season. In the long term, however, these effects are expected to decline

(Cao and Woodward, 1998; Schimel, 2007). Influences of precipitation on SOC change are more complex. Precipitation events can promote soil respiration in dry area but can hinder respiration in humid soil (Chen *et al.*, 2004). Meanwhile, the increment in precipitation can enhance the photosynthetic rate of vegetation in temperate grasslands and thus result in the accumulation of SOC stock (Yang *et al.*, 2010a). Burke *et al.* (1989) suggested that SOCD in temperate grasslands increased with precipitation in the Great Plain of the United States. Zhao *et al.* (2013) found that NPP in the northern semi-arid region of China dominated by steppe decreased with future climate changes, and to a great extent, pulse rainfall can contribute to this trend. Therefore, assessing the effects of climate change on SOC requires an integrated consideration on the sensitivities of plant production and SOC decomposition to climate change.

On the basis of the simulated result, heterotrophic respiration is not detected to increase significantly, even slightly decrease in the period after 2040 under the A1B scenario (Figure 11). Meanwhile, NPP may decline obviously with warming climate, and this trend is similar to SOCD. Therefore, we inferred that SOCD decrement under climate change scenarios can be mainly decided by declining NPP (Figure 11). In addition, the different trends of SOCD

Figure 9. SOCD spatial variation in middle term (2041–2070), compared with baseline term, under A1B, A2, and B2 scenarios.

Figure 10. SOCD spatial variation in long term (2071–2100), compared with baseline term, under A1B, A2, and B2 scenarios.

before and after 2025 are associated with the variations in NPP. Before 2025, NPP is projected to be a slightly declining trend. After 2025, NPP in Inner Mongolia shows a significantly declining trend, which can cause SOCD decrement in the same period under climate change scenarios. The changes in NPP can be explained by the spatial distribution of vegetation. The dominant vegetation type in Inner Mongolia is temperate grassland, which accounts for 67% of the total land area, distributed in Central and Western Inner Mongolia. Studies suggest that water availability is a key limiting factor that controls grassland growth (Niu, 2001; Piao *et al.*, 2006). Moreover, a significant increase in temperature may reduce water effectiveness through enhancing evapotranspiration, which may lead to a negative consequence for grassland NPP, further affecting litter input into soil. In Eastern Inner Mongolia, wherein forests are distributed, a slight increase in temperature can benefit forest NPP because of the low temperature effects. By contrast, a significant increase in temperature can weaken water availability through thawing frozen soil and increasing evapotranspiration.

Regional differences are evident in the simulated responses of SOC to climate change. We found that the change rate of SOCD is large in Western Inner Mongolia, where vegetation is sensitive to climate change, i.e. a small change in climate may result in a large impact on NPP, thus affecting the magnitude of litter input to soil. However, the change in the percentage of SOCD in

Eastern Inner Mongolia is relatively small because of the strong buffer capacity to climate change of this region. However, a small SOCD change may release more CO₂ in the air because of the large initial SOCD value. On the basis of soil survey data, Yang *et al.* (2010a, 2010b) found that the change in the percentage of SOCD is large and small in areas with low and high SOCD, respectively. Their result is consistent with that of this study.

Soil contains the largest carbon stock in the terrestrial ecosystem and is sensitive to climate change. A considerable number of studies have been conducted in the site scale compared with works in the regional scale (Xie *et al.*, 2007; Yang *et al.*, 2010a, 2010b). This study used a dynamic vegetation model to simulate the spatial change in SOC in Inner Mongolia under three climate scenarios and analysed the future change direction and degree of SOCD. However, this study is based on natural vegetation and does not consider the influence of human activities. Human activities, particularly land use changes, may exert a large impact on the SOCD (Bondeau *et al.*, 2007). The influence is considerable in Eastern Inner Mongolia, which is a typical agropastoral transitional zone, followed by Western Inner Mongolia, wherein recent mineral resource development has caused land use changes (Liu *et al.*, 2009). These effects can result in a number of uncertainties in the simulation.

In this study, the response of SOCD in Inner Mongolia to climate change was simulated by using a dynamic

Figure 11. Inter-annual variation of NPP (a) and heterotrophic respiration (b) in Inner Mongolia from 1991 to 2100 under climate change scenarios.

vegetation model. Compared with other ecological models, physiological and ecological processes, such as the instantaneous response of stomata conductance to climate change and the effects of extreme climatic events to plant growth, were considered in the dynamic vegetation model. The recent effects of extreme climatic events to ecological process have received considerable attention because such events may induce the death of plants and change the process and direction of succession, thus causing changes in the litter input. The algorithm of the dynamic vegetation model is more complex than that in the comprehensive ecological processes models (Zaehle *et al.*, 2005). To improve operation efficiency, some processes were simplified. For example, the nitrogen influences in the SOC simulation were not explicitly considered. However, the model assumed that nitrogen was fully provided; this parameter is not consistent with the actual data. Nitrogen deposition caused by fossil fuel combustion, which may affect the decomposition and conversion of SOC (Hungate *et al.*, 2003), may be added in the future model.

5. Conclusion

The SOCD in Inner Mongolia will be impaired by future climate change. However, the spatial pattern of SOCD, which decreases from the northeast to the southwest, will not be altered by climate change. In the near term, SOCD will be slightly affected by climate change, and areas showing SOCD increments and decrements will be almost equal. In the middle term, the negative influence of climate change will be enhanced, and areas with SOCD decrements will increase. In the long term, the negative effect of climate change will be prominent, and the SOCD in most eco-geographical regions will undergo large decrements. The change in the percentage of SOCD will be large in Western Inner Mongolia.

Acknowledgements

This study was funded by the National Program on Key Basic Research Project of China (No. 2011CB403206),

National Scientific Technical Supporting Programs during the 12th Five-year Plan of China (No. 2012BAC19B04). Authors are grateful to Prof. Yinlong Xu from the Institute of Environment and Sustainable Development in Agriculture, Chinese Academy of Agriculture Sciences, for providing climate scenario data.

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