Catena 116 (2014) 95–104

Contents lists available at ScienceDirect

Catena

journal homepage: www.elsevier.com/locate/catena

Soil organic carbon and influencing factors in different landscapes in an arid region of northwestern China

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ARTICLE INFO

Article history:
Received 10 April 2013
Received in revised form 21 December 2013
Accepted 24 December 2013

Keywords:
Spatial variability
Pedotransfer function
Texture
Desert
Cropland
Wetland

ABSTRACT

Knowledge of the spatial pattern of soil organic carbon (SOC) and the factors influencing it in various landscapes is essential for understanding carbon cycles. An arid region with an area of 100 km^2 in northwestern China consisted of desert, cropland and wetland was investigated. The vertical patterns of SOC density in the three different landscapes and the horizontal distribution of SOC density in the study area were evaluated. The differences in SOC density among different landscapes and soil layers were analyzed, and the primary factors influencing SOC density were determined. The density of SOC was low and remained homogeneous in the profiles of desert soil. The vertical distributions of SOC density in cropland and wetland were well described by logarithmic functions ($R^2 = 0.97$ and 0.92, respectively, $P < 0.001$). Geostatistical analysis showed that SOC density presented moderate spatial variability and strong spatial dependence across all depths. Wetland and desert were easily recognized by the highest and lowest SOC densities in the study area, respectively. The densities of SOC in the 3-m profiles were 59.35, 149.6 and 174.4 Mg ha^{-1} for desert, cropland and wetland, respectively. The SOC in the 1–3 m layer represented 67.0, 52.7 and 58.0% of the total SOC stored in the 0–3 m profiles of desert, cropland and wetland, respectively. Clay and silt particles were the major determinant of SOC in the study area. The variability in SOC density explained by clay + silt content increased with depth ranging from 46.0 to 82.2% in desert and from 45.3 to 76.7% in cropland. The variability in SOC density accounted for by clay + silt content decreased from 52.2% in the 0–0.3 m layer to 43.3% in the 0–1 m layer of wetland. The remaining SOC density variability could be attributed to factors not included in this study, such as geography, vegetation and the degree of erosion. Errors in the measurement of SOC concentration and the distribution of soil-particle size, however, may introduce uncertainty in the determination of soil bulk density and thus the estimation of SOC density. The concentration of SOC in the 0–0.3 m layer increased by 196.3% after the reclamation of native desert less than 40 years ago and decreased by 5.3% after the cultivation of wetland as cropland for less than 30 years. Short-term cultivation is insufficient to significantly alter SOC concentration in the deeper layers of desert and wetland soils. The results of this study may be of further use in optimizing strategies for the protection of wetland, ecological restoration of desertified land and the sustainable management of cropland in arid regions of northwestern China.

1. Introduction

Soil is generally recognized as a major reservoir in global carbon cycling (Batjes, 1996; Lal, 2008). Much more organic carbon is sequestered in soils than in the atmosphere and vegetation combined (Grace, 2004). Changes in soil organic carbon (SOC) are responsible for variation in the physical, chemical and biological properties of soil and influence not only crop productivity and soil fertility (Maia et al., 2010), but also regional and/or global carbon cycles (Post and Kwon, 2000). The sequestration of SOC is therefore of great concern in research on carbon cycling. Understanding the spatial variability of SOC and the primary factors influencing it are essential for evaluating the functioning of soil and understanding the process of carbon sequestration in soil.

The SOC is controlled by various natural and anthropogenic factors. For example, climate change influences the mineralization of soil organic matter (SOM) and the flux of carbon from soil to the atmosphere (Rustad and Fernandez, 1998). Topographical factors (e.g. elevation, slope and aspect) determine the production and decomposition of plant litter by controlling the soil–water balance and thus impact SOM levels (Griffiths et al., 2009). The soil–water balance may be a critical factor in determining SOC content because it integrates climate, pedological properties, topographical features and management strategies (Grigal and Ohmann, 1992). Soil–water stress may decrease, and high temperatures enhance, the decomposition of SOM (Norton et al.,

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0341-8162/$ – see front matter © 2014 Published by Elsevier B.V.
http://dx.doi.org/10.1016/j.catena.2013.12.014

2008). Soil texture distinctly influences SOC because the silt and clay particles and micro-aggregates can protect SOM from decomposition (Zinn et al., 2005). Furthermore, the roles of clay and silt particles in the availability of water (Schimel and Parton, 1986) and in plant productivity (Hontoria et al., 1999) also impact SOM. Changes in land use strongly impact soil fertility and the variation in SOC storage (Karhu et al., 2011; Z. Wang et al., 2012), especially in arid and semiarid lands where pools of SOC are susceptible to changes in land use and climate (Su et al., 2009; West et al., 1994). Management strategies such as conservation tillage, crop rotation, residue return, balanced fertilization and elimination of bare fallow, can increase SOC concentration or stocks (Mishra et al., 2010; Yang et al., 2012). Soil erosion may represent a source or sink of carbon due to the spatial variability resulting from the translocation and burial of carbon (Lal, 2003; Quinton et al., 2010). In addition to the individual influences, the combined effects of environmental and anthropogenic factors on SOC have also been extensively evaluated. For example, increased air temperature and land use change in the Himalayan region of India resulted in a decline of SOC content by 0.3% from 1978 to 2004 (Martin et al., 2010). Elevation, slope, clay and water contents of soil jointly explained 70.3% of the variability in SOC in the upstream watershed of Miyun Reservoir, North China (S.F. Wang et al., 2012).

Accurately evaluating SOC is difficult due to the complexity of site properties. The SOC particularly varies in areas with complex landforms and frequent anthropogenic disturbances. The spatial distribution of SOC and the factors influencing it have usually been studied in small areas, e.g. tens of square meters or a few square kilometers (Don et al., 2007; Han et al., 2010; Rossi et al., 2009; Wang et al., 2011). Direct measurements at regional scales are usually made at shallow soil depths such as the top 0.3–1 m (Batjes, 2006; Zinn et al., 2005). Half of the carbon released in pastures after deforestation, however, is from layers below 1 m (Batjes and Sombroek, 1997). In comparison, Batjes (1996) reported that large amounts of organic carbon are stored in the 1–2 m layer, and much of the carbon at depth is stable and does not contribute much to CO_2 emission. Organic carbon within deeper soils in regions where soils are deep and where subsoils contain considerable amounts of organic carbon should be considered.

Even though arid regions have generally been regarded as potential carbon sinks (Grünzweig et al., 2003), little information is available on SOC in arid inland areas of northwestern China, where pools of SOC are susceptible to wind erosion and to changes in land use and management practices. In the middle reaches of the Heihe River, native desert has been transitioned to irrigated cropland in recent decades to meet the food and economic demands of increasing populations (Li et al., 2006). Excessive exploitation and unreasonable management have led to soil degradation, such as severe erosion and salinization. Ecological restoration has been implemented since 1975, and some desertified land in marginal oases has been reclaimed to irrigated cropland (Su et al., 2007). In the southern part of this region, economic demands have also driven the exploitation of meadow wetland for irrigated cropland. Cropland reclaimed from native desert has been cultivated for less than 40 years. Cropland in the old oases on the banks of the Heihe River and in the southern part of the study area has been continuously cultivated for more than 100 years. Cropland exploited from meadow wetland has been cultivated for less than 30 years. Information on the period of cultivation of croplands has been collected by reviewing the literature (Su et al., 2010) and by interviewing peasants and village elders. Changes in land use substantially affect the level of SOC, and may provide information for the role of landscapes in carbon cycles. The detailed objectives of this study were: (1) to investigate the vertical distribution of SOC density in three different landscapes and its horizontal distribution in the study area and (2) to analyze the differences in SOC density among different landscapes and soil layers and to determine the major factors influencing SOC in an arid region of northwestern China.

2. Materials and methods

2.1. Study area

This study was conducted in the vicinity of the Linze Inland River Basin Research Station, Chinese Ecosystem Research Network, located in Linze County, Gansu Province, China. The area has a continental arid climate. The mean annual precipitation is 117 mm, approximately 60% of the precipitation occurs from July to September. The mean annual temperature is 7.6 °C, and the mean annual potential evaporation is 2390 mm. The mean annual wind speed is 3.2 m s^{-1} and annual gale days (wind speed varies from 17.2 to 20.7 m s^{-1}) reach to 15 or more (Su et al., 2007), and the sand-drifting mainly occurs from March to May.

Our study covered an area of 100 km^2 (20 × 5 km), with the length from north to south and the width from east to west (39°12′30″–39°23′28″N, 100°05′32″–100°10′01″E, 1372–1417 m a.s.l.). The northern part of the study area includes the southern margin of the Badain Jaran Desert. The Heihe River flows across the area from east to west. Zonal soil in the northern marginal oasis is an Aridisol derived from diluvial–alluvial materials (Su et al., 2010). Entisols form after the long-term encroachment of drift sand from the Badain Jaran Desert and the deposition of aeolian sand. In the old oases in the central and southern parts of the study area, Siltigi-Orthic Anthrosols develop under long-term irrigation from sediment-rich water, fertilization and cultivation (Su et al., 2009). Inceptisols develop in meadow wetland in the southwestern part of the study area with severe salinization occurring at the soil surface. The desert vegetation consists of *Halaxylon ammodendron* (C. A. Mey.) Bunge, *Calligonum mongolicum* Turcz., *Tamarix chinensis* Lour., *Nitraria sphaerocarpa* Maxim and *Reaumuria soongorica* (Pall.) Maxim. The predominant species in wetland are Common Reed (*Phragmites australis* (Cav.) Trin. ex Steud.) and Common Leymus (*Leymus secalinus* (Georgi) Tzvel.), interspersed with *Achnatherum splendens* (Trin.) Nevski, *Kalidium foliatum* (Pall.) Moq. and *Nitraria tangutorum* Bobr. The main crops of the irrigated cropland are maize (*Zea mays* L.) for seeds, spring wheat (*Triticum aestivumLinn.*), tomatoes (*Solanum lycopersicum*) and sugar beets (*Beta vulgaris*). Fig. 1 presents the location of the study area in China, the land use and the sampling locations. The photographs of desert, cropland and wetland are representative of the landscapes analyzed in this study.

2.2. Field investigation and sampling design

A regular grid of 126 sampling locations 1 × 1 km in size was originally designed, but six locations could not be sampled due to the presence of the river channel, reservoirs, an elevated water table, roads or buildings. The remaining 120 locations (56, 43 and 21 locations for desert, cropland and wetland, respectively) were sampled to a depth of 1 m. Of the 120 locations, 112 (56, 42 and 14 for desert, cropland, and wetland, respectively) and 102 (52 for desert, 38 for cropland, and 12 for wetland) locations were sampled to depths of 2 and 3 m, respectively. The position of each sample location was recorded with a hand-held differential GPS receiver to an accuracy of 3–5 m. Disturbed soil samples were collected with a hand auger 5 cm in diameter every 10 cm from the 0–1 m layer and every 20 cm from the 1–3 m layer by pooling two 10-cm subsamples. Soil samples for the 0–10, 10–20 and 20–30 cm layers were taken with five subsamples randomly collected from a 5 × 5 m plot. A total of 2285 soil samples were collected and sealed in air-tight bags for measuring SOC concentration and the mechanical composition immediately after transport to the laboratory. Forty of the 120 locations in the study area were selected as typical representatives, 16 of which were from desert, 18 from cropland and six from wetland locations. These locations were in the centers of each landscape where the topography and vegetation were typical. Profile was dug to a depth of 1.2 m at each of the representative locations. Eleven undisturbed soil samples (10 for the 0–1 m layer and one for

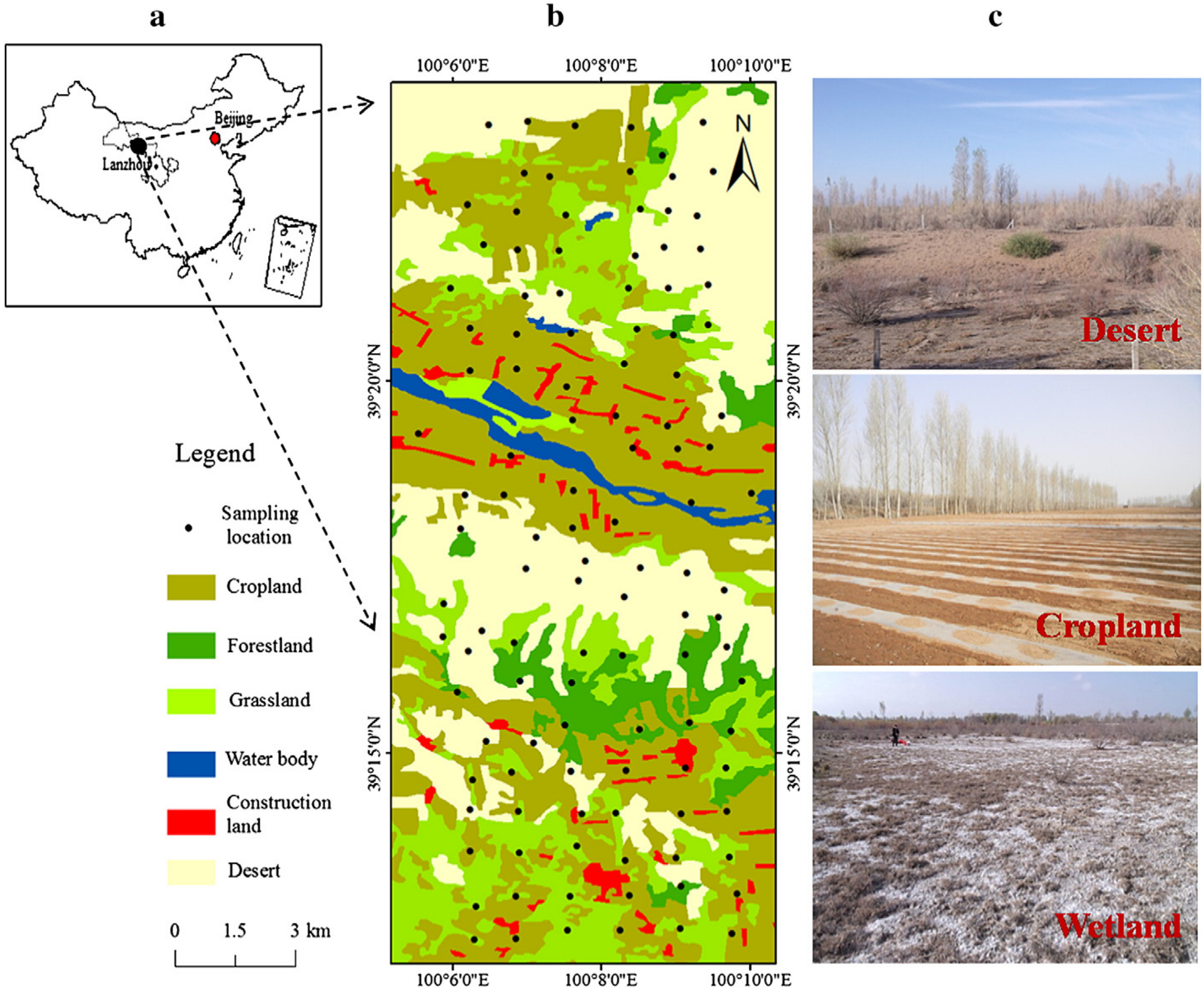

Fig. 1. The location of the study area in China (a), the sampling locations in the 100 km^2 study area (b) and the representative photographs of the three landscapes in the study area (c).

the 1.0–1.2 m layer) were collected from each profile using stainless-steel cutting ring (5.0 cm in length by 5.0 cm in diameter) for measuring soil bulk density. All samples were collected from 20 April to 10 May 2011.

2.3. Laboratory analysis

Each soil core was oven dried at 105 °C for 48 h and the dry weight of the soil was used to calculate the bulk density. Disturbed soil samples were air-dried and divided into two subsamples. One was sieved through a 2-mm mesh for particle-size analysis by laser diffraction using a Mastersizer 2000 analyzer (Malvern Instruments, Malvern, England). The other was passed through a 0.25-mm mesh for determining SOC concentration by dichromate oxidation (Walkley and Black, 1934).

As an important physical property of soil, bulk density is indispensable for the assessment of SOC density. The measurement of soil bulk density for deeper layers is usually labor intensive and time consuming. Pedotransfer functions based on easily obtained soil properties offer an alternative for estimating soil bulk density. Previous studies developed various pedotransfer functions using SOM content (De Vos et al., 2005), SOC content (Meersmans et al., 2009a) or the combination of SOC content and soil texture (Kaur et al., 2002) as predictor variables in soils where OM has a dominant effect on soil bulk density. Texture or other properties may significantly contribute to bulk density in soils where OM is a minor component. An equation specific for each range of soils relevant to a particular study is recommended for estimating bulk density with high accuracy and precision, rather than relying on generalized pedotransfer functions (Harrison and Bocock, 1981). Of the 40 representative locations in this study, 16, 18 and six belonged to desert, cropland and wetland, respectively. Correlations among the distribution of soil-particle size, SOC concentration and bulk density in the 1.2-m profiles of each landscape were analyzed. Stepwise multiple linear regressions using SOC concentration and clay, silt and sand contents as independent variables and bulk density as a dependent variable were conducted for desert, cropland and wetland, respectively. Once the pedotransfer function for each landscape was satisfactory, bulk density in each layer was estimated, and SOC density for a profile with n layers was calculated (Batjes, 1996):

$$Density\ \ of\ \ SOC_h = \sum_{i=1}^{n} \frac{SOC_{Ci} \times BD_i \times D_i \times (1-S_i)}{10} \tag{1}$$

where *density of* SOC_h is the total amount of SOC over depth h per unit area (Mg ha^{-1}); n is the number of soil layers considered; SOC_{Ci}, BD_i and D_i represent SOC concentration (g kg^{-1}), bulk density (g cm^{-3}) and thickness (cm) of the ith layer, respectively, and S_i is the volume of the fraction of coarse fragments >2 mm in the ith layer. S_i is negligible due to the very low contents of coarse fragments in the soils of the study area (Zhang et al., 2012).

2.4. Statistical and geostatistical analyses

The Kolmogorov–Smirnov (K–S) test examined the normality of the distribution of soil properties. The raw dataset of SOC density in each layer of the entire study area or of individual landscapes was skewed. The data were normalized by logarithmic (log10) transformation. Latitude, longitude and elevation were normally distributed. The bulk density, sand, clay and silt contents in the different layers of each landscape were all normally distributed. One-way analyses of variance (ANOVA) were performed on SOC concentration, SOC density and clay + silt content among landscapes or soil layers. Pearson correlation coefficients determined the strength of possible relationships among the distribution of soil-particle size, SOC concentration and bulk density at representative locations of each landscape and between SOC density and the site variables at all sampling locations. Statistical analyses were performed using SPSS software (version 17.0, SPSSInc.) The spatial structure of SOC density was analyzed with geostatistical software (GS+, version 7.0, Gamma Design Software). Kriging was used for interpolation and for producing maps with ArcGIS for desktop (version 9.3, ESRI).

3. Results

3.1. Estimation of SOC density

Wetland had the highest clay and silt contents, the highest SOC concentration and the lowest sand content in the profiles of the representative locations and at all locations (Table 1). Desert had the lowest SOC concentration and clay and silt contents and the highest sand content. The representative wetland locations had the lowest soil bulk density, while the desert locations had the highest (Table 1). Bulk density, SOC concentration and clay and silt contents were positively correlated, and sand content and bulk density were negatively correlated, at the desert representative locations. In contrast, SOC concentration and clay and silt contents were negatively correlated with bulk density at the cropland and wetland representative locations (Table 2). Application of the stepwise regression analysis to the representative locations allowed the inclusion of different variables in the regression models for the three landscapes (Table 3). For desert, the 2nd-order polynomial regression model used clay, silt and sand combined, and clay and SOC concentration combined as predictor variables. For cropland, the 3rd-order polynomial model confirmed sand, clay and sand combined, and SOC concentration and clay combined as predictor variables. For wetland, only SOC concentration and silt and SOC concentration combined remained as predictor variables in the 2nd-order polynomial model. Adjusted R^2 values of the regression models revealed that the 2nd-order, 3rd-order and 2nd-order polynomial regression models accounted for 34, 41 and 68% of the variability in soil bulk density for desert, cropland and wetland, respectively. The unstandardized coefficients for the predictor variables were all significant ($P < 0.001$, Table 3). Furthermore, the standard errors of estimates for the polynomial regression models were low (Table 3). The polynomial regression models were regarded as representative of the associations among soil bulk density, SOC concentration and the distribution of particle size. This concurs with the result that polynomial regression models reflected the association between bulk density and basic soil properties better than general linear model (Kaur et al., 2002). The densities of SOC in the different layers of each landscape were thereby estimated.

Table 2
Correlation coefficients between soil bulk density and soil organic carbon concentration (SOC_c), clay, silt and sand contents at the 40 representative locations for the three landscapes in the study area.

Variable [a]		Clay (%)	Silt	Sand	SOC_c (g kg^{-1})
Bulk density (g cm^{-3})	Desert	0.338** [b]	0.214**	−0.330**	0.170*
	Cropland	−0.132	−0.308**	0.345**	−0.418**
	Wetland	−0.280*	−0.611**	0.498**	−0.786**

[a] Bulk density, clay, silt and sand contents and soil organic carbon concentration for desert, cropland and wetland are all normally distributed ($P > 0.05$).
[b] ** and *, the correlations are significant at the 0.01 and 0.05 levels, respectively.

3.2. Vertical distribution of SOC density

The vertical distributions of SOC density in the 3-m profiles at 20-cm intervals for the three landscapes are shown in Fig. 2. The densities of SOC in the profiles ranged from 3.64 to 4.48, 6.70 to 20.31 and 8.27 to 19.57 Mg ha^{-1} for desert, cropland and wetland, respectively. The density of SOC was relatively high in the 0–20 cm layer of desert soil (4.48 Mg ha^{-1}) and was low and remained homogeneous (ranging from 3.64 to 4.13 Mg ha^{-1}) in deeper layers. The densities of SOC in cropland and wetland decreased logarithmically with increasing depth (R^2 = 0.97 and 0.92, respectively, $P < 0.001$, Fig. 2).

3.3. Spatial variability and distribution of SOC density

The coefficients of variation for SOC density in different layers indicated moderate spatial variability, which decreased with depth, ranging from 77.0 (0–0.3 m) to 60.3% (2–3 m) in the study area. The SOC in surface layers was more susceptible to changes in climate, vegetation, land use and management practices, and the interspersion of different landscapes in the study area created a complex landscape pattern. The spatial heterogeneity of the site properties contributed remarkably to SOC density variability in the shallow layers.

The spatial variability of SOC density was determined by geostatistical analysis. Anisotropy was not detected in the directional semivariograms of SOC density because the anisotropic ratios (ranging from 1.04 to 1.80, Table 4) were less than 2.5 (Wang et al., 2010). The isotropic spherical models fit the experimental semivariograms best, with the highest

Table 1
The mean values of soil properties at the 40 locations selected as representatives of the three landscapes and at all 120 locations in the study area.

Variable	Representative locations			All locations		
	Desert	Cropland	Wetland	Desert	Cropland	Wetland
n [a]	16(176)	18(198)	6(66)	56(1100)	43(833)	21(354)
Clay (%) [b]	8.27 ± 10.3	29.2 ± 11.1	30.8 ± 9.52	12.5 ± 14.5	26.9 ± 13.8	31.5 ± 12.1
Silt (%)	10.1 ± 11.2	41.1 ± 13.2	47.7 ± 8.32	15.7 ± 16.4	37.6 ± 15.8	45.0 ± 12.1
Sand (%)	81.6 ± 20.8	29.8 ± 22.1	21.4 ± 15.6	71.8 ± 29.8	35.5 ± 27.7	23.5 ± 21.7
SOC_c (g kg^{-1}) [c]	1.09 ± 0.67	5.09 ± 3.26	5.66 ± 3.05	1.22 ± 0.79	3.66 ± 2.56	4.38 ± 2.67
BD (g cm^{-3})	1.59 ± 0.06	1.55 ± 0.15	1.49 ± 0.17	–	–	–

[a] Data out of the brackets is the number of locations for each landscape, and data in the brackets is the sample number in the 1.2 m profiles for each landscape among the 40 representative locations and in the 3 m profiles for each landscape among the 120 locations.
[b] Mean ± standard deviation.
[c] SOC_c, soil organic carbon concentration. BD, bulk density.

Table 3
The stepwise multiple linear regression of soil bulk density with soil organic carbon concentration, clay, silt and sand contents at the 40 representative locations for the three landscapes in the study area.

Landscape	n [a]	Predictor variable [b]	Unstandardized coefficient	Adjusted R^2	S.E. of estimate	P
Desert	16(176)	Constant	1.58^{***} [c]	0.34	0.05	<0.001
		Clay	$9.18 \times 10^{-3***}$			
		Silt*Sand	$5.18 \times 10^{-5***}$			
		Clay*SOC_c concentration	$2.14 \times 10^{-3***}$			
Cropland	18(198)	Constant	1.36^{***}	0.41	0.12	<0.001
		$Clay^2$*Sand	$1.21 \times 10^{-5***}$			
		$Sand^3$	$3.61 \times 10^{-7***}$			
		SOC_c^2*Clay	$3.00 \times 10^{-5***}$			
Wetland	6(66)	Constant	1.80^{***}	0.68	0.10	<0.001
		SOC_c^2	$3.00 \times 10^{-3***}$			
		Silt*SOC_c	$1.59 \times 10^{-3***}$			

[a] The number of representative locations for each landscape (data out of the brackets) and the number of samples in the 1.2 m profiles at representative locations of each landscape (data in the brackets).
[b] Clay, clay content (%); Silt, silt content (%); Sand, sand content (%); SOC_c, soil organic carbon concentration (g kg^{-1}).
[c] ***, the coefficient is significant at the 0.001 level.

values of R^2 and the lowest residual sum of squares (Table 4). The low nugget values indicated a weak field or random variability, and low nugget-to-sill ratios (5.51–17.8%) suggested a strong spatial dependence of SOC density in the four soil layers (Cambardella et al., 1994). These results indicated that SOC density may be controlled by the intrinsic variations in soil properties, such as parent material, soil texture and mineralogy (Cambardella et al., 1994; Wang et al., 2010). The areas of wetland were derived from alluvial materials and exhibited high levels of pedogenic development and high fractions of fine particles. Cropland presented moderate levels of pedogenic development and loamy parental materials. Desert soils were developed from aeolian sand and consisted mainly of well-sorted medium-fine sand (Su et al., 2010). The increasing nugget-to-sill ratios indicated that the spatial dependence of SOC density decreased with depth. The 2–3 m layer had the largest nugget and the smallest sill, indicating the highest random variation and the lowest spatial dependence of SOC density. Range reflects the largest separation distance within which soil property is spatially related. The ranges of SOC density in the four soil layers varied from 5.24 to 5.65 km, and were larger than the sampling interval (1 km). The sampling density was thus sufficiently intensive to obtain information on the spatial variability of SOC density. The similar ranges of SOC density among the different soil layers indicated that the zones of influence of the site variables on SOC were relatively uniform and were not depth dependent.

The distribution maps of SOC density in the 0–1, 1–2 and 2–3 m layers are shown in Fig. 3. The 0–1 m layer stored more organic carbon in each landscape. Wetland in the southwestern part of the study area

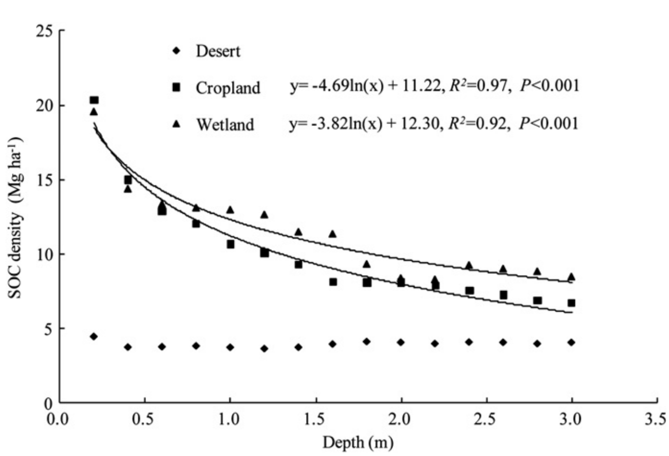

Fig. 2. The vertical distribution of soil organic carbon density for desert, cropland and wetland.

and cropland adjoining the Heihe River could be easily recognized by their high SOC densities in the three layers. Desert in the northern and central parts was characterized by low SOC densities.

3.4. The variability of SOC density

3.4.1. Variability of SOC density among landscapes

Landscapes significantly affect the patterns of vertical distribution of SOC density. The density of SOC was highest in wetland and lowest in desert across all depths except in the 0–0.3 m layer where SOC density in cropland was the highest (Fig. 4). Differences in SOC density between wetland and cropland were not statistically significant in the 0–0.3, 0–1 and 2–3 m layers ($P > 0.05$, Fig. 4). The density of SOC in the 1–2 m layer was significantly higher in wetland than in cropland ($P < 0.05$, Fig. 4). The densities of SOC in wetland and cropland were significantly higher than in desert across all depths ($P < 0.05$, Fig. 4). The densities of SOC in the 0–0.3 m layer of desert, cropland and wetland were 6.39, 28.32 and 27.14 Mg ha^{-1}, representing 32.7, 40.0 and 37.0% of SOC densities of the 0–1 m layer, respectively. The uppermost 10 cm of desert, cropland and wetland contained 38.7, 38.1 and 39.5% of SOC in the 0–0.3 m soil, respectively. The densities of SOC in the 0–1 m layer were 19.57, 70.83 and 73.33 Mg ha^{-1} for desert, cropland and wetland, respectively. The associated values in the 1–2 and 2–3 m layers were 19.55 and 20.23 Mg ha^{-1} for desert, 42.60 and 36.19 Mg ha^{-1} for cropland and 57.35 and 43.76 Mg ha^{-1} for wetland. The densities of SOC in the 0–0.3, 0–1, 1–2 and 2–3 m layers were an average of 4.3, 3.7, 2.6 and 2.0 times higher, respectively, in cropland and wetland than in desert. The density of SOC in the 0–3 m layer of wetland was 174.4 Mg ha^{-1}, which was 2.9 and 1.2 times of that of desert (59.35 Mg ha^{-1}) and cropland (149.6 Mg ha^{-1}), respectively. These results reflect the potential importance of wetland on the sequestration of SOC in arid regions. The SOC in the 1–3 m layer of desert, cropland and wetland contributed 67.0, 52.7 and 58.0% of the total SOC stored in the 3-m profiles, respectively. The proportional distribution (interval/reference depth) was deepest in desert and shallowest in wetland.

3.4.2. Variability of SOC density among soil layers

The densities of SOC of desert ranged from 19.57 to 20.23 Mg ha^{-1} in the 0–1, 1–2 and 2–3 m layers, which were an average of 3.1-fold of that in the 0–0.3 m layer and 33.0, 32.9 and 34.1% of those in the 0–3 m layer, respectively. The lack of significant differences ($P > 0.05$, Fig. 4) in SOC density among the three layers reflects the homogeneous distribution of low SOC in the desert soil (Fig. 2). The densities of SOC in the 0–1, 1–2 and 2–3 m layers of cropland were 70.83, 42.60 and 36.19 Mg ha^{-1}, respectively, representing 47.3, 28.5 and 24.2% of SOC density in the 0–3 m profile, respectively. The 0–1 m layer, the main activity zone of the root system and susceptible to agricultural practices,

Table 4
Geostatistical parameters for soil organic carbon density in the 0–0.3, 0–1, 1–2 and 2–3 m soil layers in the study area.

Depth (m)	n [a]	Model	Nugget (C_0) (Mg^2 ha^{-2})	Sill ($C + C_0$)	Nugget/Sill ($C_0/C + C_0$) (%)	Range (Km)	R^2	RSS [b]	Anisotropy ratio
0–0.3	120	Spherical	8.8E−03	0.16	5.51	5.24	0.71	1.1E−02	1.80
0–1	120	Spherical	8.8E−03	0.13	6.95	5.65	0.71	7.4E−03	1.69
1–2	112	Spherical	6.8E−03	0.09	7.19	5.24	0.75	2.9E−03	1.20
2–3	102	Spherical	1.3E−02	0.07	17.8	5.33	0.59	2.8E−03	1.04

[a] n, the sample number.
[b] RSS, the residual sum of squares.

stored significantly more SOC than did the deeper layers ($P < 0.05$, Fig. 4). The densities of SOC in the 0–1, 1–2 and 2–3 m layers of wetland were 73.33, 57.35 and 43.76 Mg ha^{-1}, respectively, contributing 42.0, 32.9 and 25.1% of SOC density in the 0–3 m profile, respectively. Significantly more SOC was stored in the upper layers ($P < 0.05$, Fig. 4).

4. Discussion

The homogeneously low SOC density in the profile of desert soil was consistent with the findings of Su et al. (2007), who reported a relatively low and uniform distribution of SOC concentration in the 0–1 m layer of sandy soil. Even though an exponential decline of SOC concentration with depth was usually reported (Meersmans et al., 2009a; Sleutel et al., 2003), the vertical distributions of SOC density in cropland and wetland in our study were best fit by logarithmic functions ($R^2 = 0.97$ and 0.92, respectively, $P < 0.001$, Fig. 2). The density of SOC of cropland decreased from 10.79 Mg ha^{-1} in the 0–10 cm layer to 8.01 Mg ha^{-1} in the 20–30 cm layer, which differed from the constant SOC density at tillage depth (0–30 cm) of cropland in Flanders (Meersmans et al., 2009a). This difference may be due to the distinct climates, soil types and management practices between the two regions. The 1–3 m layer of desert, cropland and wetland contained 67.0, 52.7 and 58.0% of the total SOC, respectively. The pool of SOC in deeper layers may be ascribed to seepage by water transport, soil permeability, earthworm bioturbation, leaching and vertical mixing of SOC (Don et al., 2007; Jobbágy and Jackson, 2000), the decreasing SOC turnover (Trumbore, 2000; Wei et al., 2012) and increasing root turnover with depth (Jobbágy and Jackson, 2000). The desert area in this study was characterized by larger pores and higher permeability, which contributed greatly to the deepest proportional distribution of SOC. The infiltration and percolation of irrigation water in cropland transports SOM and fine particles to the deep layers. Leaching in wetland carries salts and SOM to the deep layers where saturated water contents and high fractions of fine particles protect SOC against oxidation. The deep pools of SOC in the different landscapes raise important issues for the strategies of carbon sequestration in arid regions.

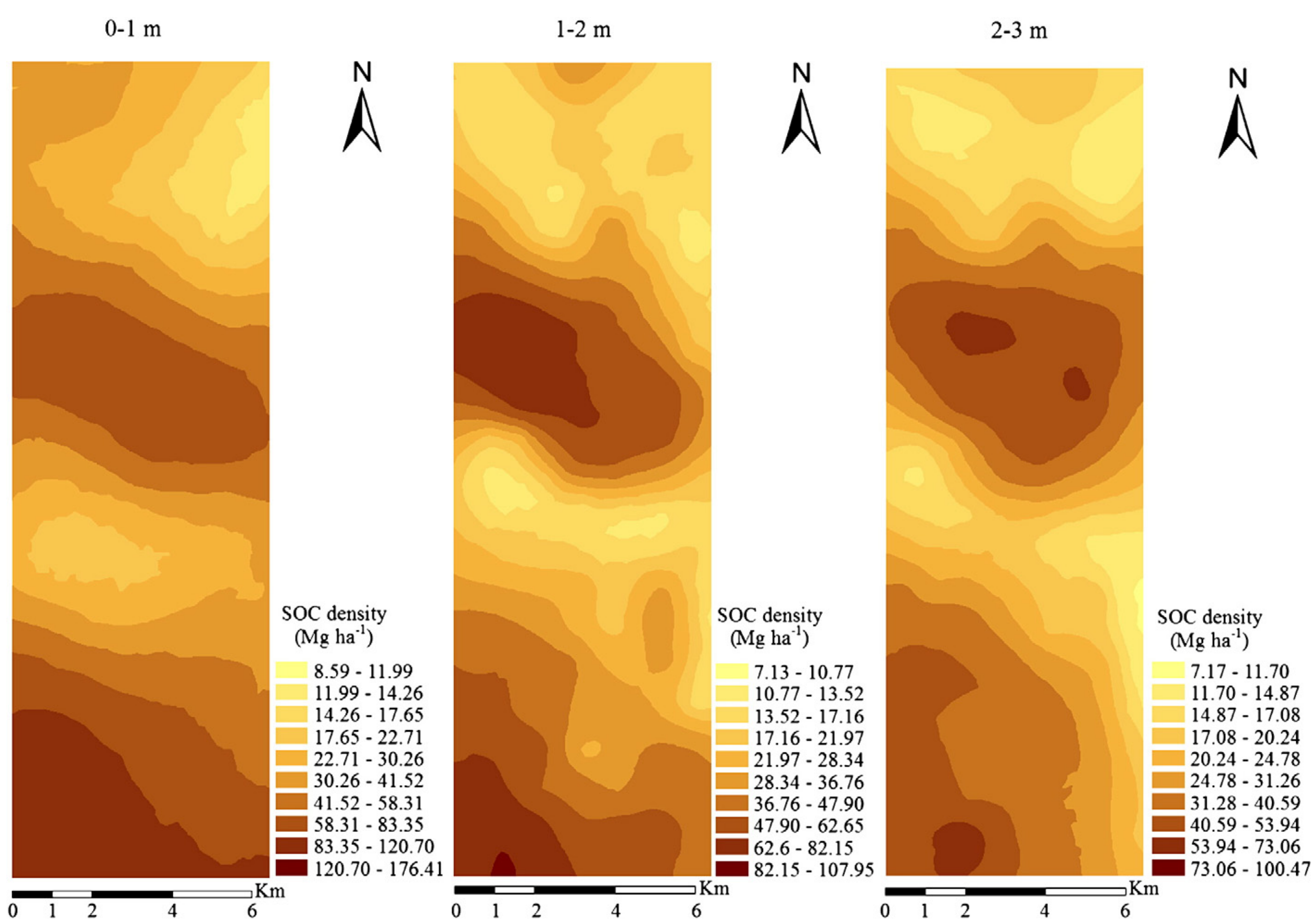

Fig. 3. The spatial distribution of soil organic carbon density in the 0–1, 1–2 and 2–3 m soil layers in the study area.

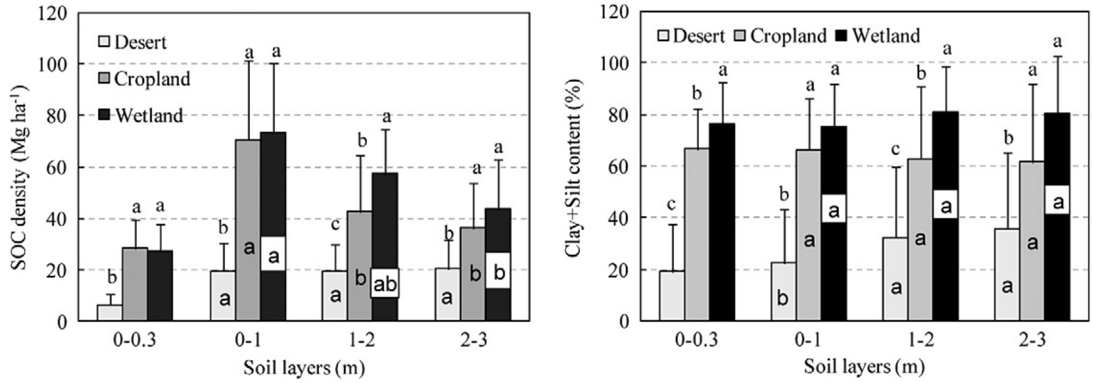

Fig. 4. Comparisons of soil organic carbon density and of clay + silt content among different landscapes and soil layers. Letters above the bars are for comparing different landscapes for a certain soil layer, and letters in the bars are for comparing different soil layers for a certain landscape. Different lowercase letters indicate significant differences ($P < 0.05$). The error bars are one standard deviation of the means.

The SOC is affected by various properties. The density of SOC was not correlated with latitude, longitude and elevation in the three landscapes ($P > 0.05$, Table 5). The levels of explanation of SOC density variability by latitude, longitude and elevation are random in this study. Latitude and longitude varied little in this study and accounted individually or in combination for 5.20, 2.60 and 13.2% of SOC density variability in the 0–0.3 and 0–1 m layers of desert (Table 6) and the 0–1 m layer of cropland (Table 7), respectively. Elevation affects SOC directly via its influence on turnover rates or indirectly via its effects on the intensity of weathering and on the formation of secondary clay minerals (Powers and Schlesinger, 2002). Temperature declines with increasing elevation, and lower temperatures would reduce the rate of decomposition of SOM, leading to the accumulation of SOC. The concentration of SOC increased by 0.75–2.1 mg g^{-1} per 100 m of increase in elevation in agricultural soils of Switzerland (Leifeld et al., 2005). Elevation in this study ranged from 1372 to 1417 m. The mean elevation of wetland (1404 m) was higher than that of desert (1387 m) and cropland (1390 m). Elevation, however, failed to account for SOC density variability, except for the 0–0.3 m layer of desert where elevation accounted for 2.9% of the variability. This lack of association may be ascribed to the narrow range of elevation and the counteracting effect of higher SOC concentration and lower bulk density at higher elevation (Powers and Schlesinger, 2002).

The density of SOC was typically high in fine-textured soils. This result was consistent with international literature (Jobbágy and Jackson, 2000; Meersmans et al., 2008; Tan et al., 2004). Clay + silt contents were significantly and positively correlated with SOC density at all depths in desert and cropland soils and in the 0–0.3 and 0–1 m layers in wetland soil ($P < 0.01$, Table 5). This result was in accordance with the report of Zinn et al. (2005). The positive effects of clay and silt particles on the accumulation of SOC are due to the stabilization of SOM by clay particles and associated iron oxides (Lützow et al., 2006) and the promotion of aggregation and physical protection of SOC from oxidation by the relatively smaller spaces in the soil (Baritz et al., 2010). Soil clay and silt particles appear to be the major determinant of SOC density in desert and cropland, especially in deeper layers. This result kept in line with Zinn et al. (2005), who reported a linear relationship between SOC concentration and clay + silt content, and found that the textural control of SOC concentration varied predictably with depth in the 0–1 m soil of Brazilian Cerrado. The variability in SOC density explained by clay + silt content increased with depth by 46.0–82.2% in desert (Table 6) and by 45.3–76.7% in cropland (Table 7). The variability in SOC density accounted for by clay + silt content, however, decreased from 52.2% in the 0–0.3 m layer to 43.3% in the 0–1 m layer of wetland soil (Table 8). The clay + silt contents of wetland and cropland were generally 3.0- and 2.5-fold higher, respectively, than that of desert across all depths. These differences were statistically significant ($P < 0.05$, Fig. 4). The wetland soils are derived from alluvial materials and exhibit high pedological development and high levels of clay + silt particles. Cropland soils present loamy parental materials and moderate pedogenic development. Agricultural practices dramatically alter the soil-forming processes vital to soil structure and fertility. Desert soils are developed from aeolian sand and consist mainly of well-sorted medium-fine sand (Su et al., 2010). The lack of fine particles due to long-term wind erosion and aeolian transportation of sand strongly impacts the loose sandy structure. The differences in the fractions of clay + silt particles between wetland and cropland were significant only in the 0–0.3 m layer ($P < 0.05$, Fig. 4). The significantly lower clay + silt content in the 0–0.3 m cropland soil could be ascribed to wind erosion and improper agricultural management. The irrigation of crops during the growing season has totally depended on groundwater in recent decades, which has decreased the input of fine particles from the silt-laden water of the Heihe River (Su et al., 2010). Infiltration and percolation also cause the loss of fine particles from surface soil. Plowing in the autumn does not return residues to the topsoil, but fine particles are blown by severe wind erosion and are deposited elsewhere in winter

Table 5
Pearson correlation coefficients between soil organic carbon density (SOC density) and site variables in different soil layers of the three landscapes in the study area.

Variable [a]	SOC_D density (Mg ha^{-1})				SOC_C density (Mg ha^{-1})				SOC_W density (Mg ha^{-1})			
	0–0.3 m	0–1 m	1–2 m	2–3.m	0–0.3 m	0–1 m	1–2 m	2–3 m	0–0.3 m	0–1 m	1–2 m	2–3 m
n	56	56	56	52	43	43	42	38	21	21	14	12
Latitude (°)	−0.16	−0.19	−0.05	−0.04	−0.52** [b]	−0.57**	−0.10	0.10	0.13	0.07	−0.32	−0.18
Longitude (°)	0.08	0.02	−0.26	−0.26	−0.11	−0.08	−0.11	−0.09	−0.32	−0.41*	−0.16	−0.23
Elevation (m)	0.27*	0.34**	0.13	0.09	0.45**	0.44**	−0.09	−0.24	−0.02	0.03	0.38	−0.01
Clay + silt (%)	0.69**	0.84**	0.84**	0.91**	0.60**	0.66**	0.79**	0.63**	0.65**	0.62**	0.14	0.40

[a] SOC_D density, SOC_C density, SOC_W density, the logarithmically transformed SOC density for desert, cropland and wetland, respectively. n, sample number. Clay + silt, clay plus silt content.
[b] **, and *, the correlations are significant at 0.01 and 0.05 levels, respectively.

Table 6
Stepwise regression of soil organic carbon density (SOC density) with selected site variables in different layers of desert soil in the study area.

Dependent variable [a]	Independent variables	Unstandardized coefficient	Adjusted R^2	S.E. of estimate	P
$SOC_{0-0.3}$ density	Constant	278.3^{*} [b]	0.54	0.15	<0.001
	Clay+silt (%)	$9.0E-03^{***}$			
	Elevation (m)	$9.0E-03^{**}$			
	Longitude (°)	-2.90^{*}			
SOC_{0-1} density	Constant	178.3^{*}	0.73	0.10	<0.001
	Clay+silt (%)	$8.0E-03^{***}$			
	Longitude (°)	-1.77^{*}			
SOC_{1-2} density	Constant	1.04^{***}	0.70	0.11	<0.001
	Clay+silt (%)	$6.0E-03^{***}$			
SOC_{2-3} density	Constant	1.02^{***}	0.82	0.09	<0.001
	Clay+silt (%)	$6.0E-03^{***}$			

[a] The density of SOC in the 0–0.3, 0–1, 1–2 and 2–3 m soil layers, respectively, and the variables were logarithmically transformed.
[b] ***, ** and *, the coefficients are significant at 0.001, 0.01 and 0.05 levels, respectively.

and early spring. On the other hand, clay + silt content did not differ significantly among the 0–1, 1–2 and 2–3 m layers in either cropland or wetland ($P > 0.05$, Fig. 4). The significantly lower clay + silt content in the 0–1 m desert soil ($P < 0.05$, Fig. 4) can be ascribed to the severe wind erosion, which removes fine particles from the topsoil and leaves behind coarse sand.

Vegetation, through patterns of above- and belowground allocation, has an important effect on SOC (Jobbágy and Jackson, 2000). Ecological restoration conducted since 1975 in the desertified land of the study area has increased the fraction of fine particles, moisture content and fertility of the soil (Su et al., 2007, 2012). Such soil amelioration benefits the allocation and growth of annual herbaceous plants, which supply most of the carbon input to the topsoil. The concentration of SOC in the 0–10 cm layer of desert soil (1.57 g kg^{-1}) was higher than the reported value in the 0–20 cm sand land (0.90 g kg^{-1}, Su et al., 2010). Ecological restoration has effectively sequestrated carbon, and desert soil has the potential to be a carbon sink. The concentration of SOC in the 0–10 cm layer of desert soil in this study, however, was still lower than that in the neighboring Gaotai County (3.1 g kg^{-1}, Li et al., 2009). The density of SOC in the 0–1 m layer of desert soil (19.6 Mg ha^{-1}) is still lower than the global average of 30 Mg ha^{-1} for the Yermosols of arid regions (Batjes and Sombroek, 1997). Marked by densely distributed meadows and interspersed salt-tolerant shrubs, wetland has relatively shallow root profiles. Roots produce highly decomposable fine roots and root exudates, which can be rapidly decomposed by microbes when exposed to more aerobic conditions, contributing to the accumulation of SOC (Thomas et al., 1996).

The groundwater regime of wetland also controls the decomposition and accumulation of SOC (Baritz et al., 2010). The level of the water table determines the occurrence depth of anaerobic conditions (Harriss et al., 1982). High water tables or poor drainage produce shallow anaerobic conditions and reduce the oxidative capacity of SOC (Meersmans et al., 2009b). Gases held in pore water and the matrix are rapidly released when water tables are low (Moore and Dalva, 1993). Even relatively small hydrological changes due to water diversions, ditching and groundwater pumping can significantly affect the level of groundwater and thus carbon cycles (Limpens et al., 2008). The SOM contents of wetland in this study were 1.38 and 1.12% in the 0–10 and 10–20 cm layers, respectively, which were lower than the values of 1.80 and 1.40%, respectively, reported by Zhao et al. (2008). These differences may be ascribed to the decline of groundwater level due to the reliance on groundwater for crop irrigation in the study area.

Changes in land use also strongly affected SOC. For cropland developed from native desert, the concentration of SOC increased by varying degrees in different layers. Cultivation significantly increased the fractions of fine particles and SOC concentration in the surface soil ($P < 0.05$, Fig. 5a, c). An increase in SOC concentration and clay + silt contents in surface soil has also been observed after the conversion of native desert to irrigated cropland in the northern part of this study area (Su et al., 2010) and in Gaotai County (Li et al., 2009). After conversion, the clay + silt contents in reclaimed desert increased significantly by 1.88-, 2.11- and 1.43-fold in the 0–0.1, 0.1–0.3 and 0–1 m layers, respectively ($P < 0.05$, Fig. 5a). The concentration of SOC increased significantly by 1.97-, 1.96- and 1.05-fold in the 0–0.1, 0.1–0.3 and 0–1 m layers, respectively ($P < 0.05$, Fig. 5c). Agricultural management, however, was insufficient to significantly improve the soil in the deep layers. Clay + silt content increased by only 27.2 and 31.1%, and SOC concentration increased by 15.1 and 21.6%, in the 1–2 and 2–3 m layers, respectively (Fig. 5a, c). Several decades of cultivation were required for SOC concentration in reclaimed desert to reach the level of cropland in the old oases. The differences in clay + silt content between cultivated desert and cropland in the old oases were not significant in the 0–0.1, 0.1–0.3 and 0–1 m layers ($P > 0.05$, Fig. 5a). The concentration of SOC, however, was significantly lower in cultivated desert than in cropland in the old oases across all depths ($P < 0.05$, Fig. 5c). The concentrations of SOC of cropland exploited from native desert were 68.7, 67.0 and 51.6% of those of cropland in the old oases on the banks of the Heihe River in the 0–0.1, 0.1–0.3 and 0–1 m layers, respectively. The differences were considerably lower (36.1 and 48.4%) in the 1–2 and 2–3 m layers, respectively. Conventional tillage by mechanical equipment is

Table 7
Stepwise regression of soil organic carbon density (SOC density) with selected site variables in different layers of cropland soil in the study area.

Dependent variable [a]	Independent variables	Unstandardized coefficient	Adjusted R^2	S.E. of estimate	P
$SOC_{0-0.3}$ density	Constant	0.85^{***} [b]	0.45	0.15	<0.001
	Clay+silt (%)	$8.0E-03^{***}$			
SOC_{0-1} density	Constant	360.0^{***}	0.64	0.12	<0.001
	Clay+silt (%)	$6.0E-03^{***}$			
	Latitude (°)	-1.46^{***}			
	Longitude (°)	-3.01^{**}			
SOC_{1-2} density	Constant	1.04^{***}	0.77	0.13	<0.001
	Clay+silt (%)	$8.0E-03^{***}$			
SOC_{2-3} density	Constant	96.4^{*}	0.63	0.14	<0.001
	Clay+silt (%)	$6.0E-03^{***}$			

[a] The density of SOC in the 0–0.3, 0–1, 1–2 and 2–3 m soil layers, respectively, and the variables were logarithmically transformed.
[b] ***, ** and *, the coefficients are significant at 0.001, 0.01 and 0.05 levels, respectively.

Table 8
Stepwise regression of soil organic carbon density (SOC density) with selected site variables in different layers of wetland soil in the study area.

Dependent variable [a]	Independent variables	Unstandardized coefficient	Adjusted R^2	S.E. of estimate	P
$SOC_{0–0.3}$ density	Constant	0.76^{***} [b]	0.52	0.13	<0.001
	Clay+silt (%)	$8.0E-03^{***}$			
$SOC_{0–1}$ density	Constant	1.30^{***}	0.43	0.13	<0.001
	Clay+silt (%)	$7.0E-03^{*}$			
$SOC_{1–2}$ density	– [c]	–	–	–	–
$SOC_{2–3}$ density	–	–	–	–	–

[a] The density of SOC in the 0–0.3, 0–1, 1–2 and 2–3 m soil layers, respectively, and the variables were logarithmically transformed.
[b] ***, ** and *, the coefficients are significant at 0.001, 0.01 and 0.05 levels, respectively.
[c] The data do not exist.

widely used in cropland in the study area, which tends to lead to intensified and deep soil disturbance. The loss of SOM is thus accelerated via rendering the soil more susceptible to oxidation and erosion. After harvest, stalks are collected and removed for feeding livestock, no residue is returned and farmyard manure is rarely applied. Cropland is plowed with a moldboard in the autumn and is kept bare before seeding in the following April. Rational management (e.g. conservation tillage, residues return, crop rotation and intercropping, manure application and water-saving irrigation) should be introduced to minimize the loss of SOC by wind erosion and to ensure the sustainable agricultural use.

The digging of ditches in wetland before cultivation to reduce salinization has lowered the water table in the southern part of the study area. After short-term cultivation, clay + silt content and SOC concentration decreased in the soil profiles ($P > 0.05$, Fig. 5b, d). The concentrations of SOC in the 0–0.1, 0.1–0.3, 0–1, 1–2 and 2–3 m layers of the exploited wetland were 90.5, 98.9, 91.7, 66.5 and 82.4%, respectively, of those of the wetland. Agricultural practices such as irrigation and fertilization have increased SOC concentration in the surface soil as cultivation has continued. The concentrations of SOC in cropland exploited from wetland for more than 100 years were 1.25-, 1.51- and 1.31-fold of those of wetland in the 0–0.1, 0.1–0.3 and 0–1 m layers, respectively. Exploited wetland after short-term cultivation has thus acted as a carbon source, and the evolution of wetland soil toward cropland soil has been slow under current management practices. Considering the ecological influence of wetland on regional climate, water cycles and biodiversity and its role as a carbon sink, newly exploited wetland should be converted to meadow grassland, and the current natural wetland should be protected from exploitation.

5. Conclusions

Pedotransfer functions established for different landscapes effectively estimated the soil bulk density in this region. The vertical distribution of SOC density differed among landscapes. The strong spatial dependence of SOC density was associated with the soil genesis in arid regions. The fraction of clay + silt particles is the core determinant of SOC. Wetland soil derived from alluvial materials with high pedological development was characterized by the highest SOC density across all depths. Desert soil developed from aeolian sand and with a loose structure had the lowest SOC density across depths. The conversion of native desert to irrigated cropland significantly increased SOC concentration in the 0–1 m layer. The cultivation of wetland for short-term period decreased

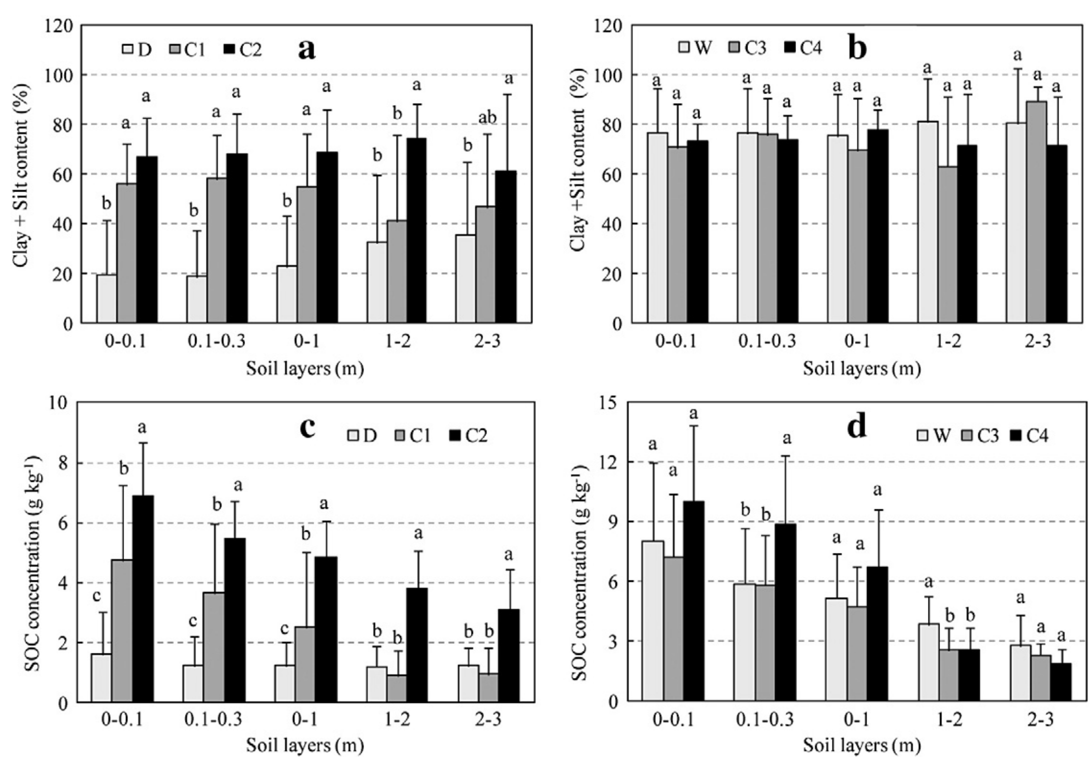

Fig. 5. Comparisons of clay + silt content (a and b) and of soil organic carbon concentration (c and d) among desert (or wetland), newly cultivated cropland and cropland cultivated continuously for more than 100 years. Different lowercase letters indicate significant differences ($P < 0.05$). The error bars are one standard deviation of the means. D, desert; C1, desert reclaimed for less than 40 years; C2, cropland cultivated for more than 100 years in the old oases; W, wetland; C3, wetland cultivated for less than 30 years; C4, cropland exploited from wetland for more than 100 years.

SOC concentration and the exploited wetland acted as a carbon source. Rational management (e.g. no-till or minimum tillage, residues return, crop rotation and intercropping, manure application and water-saving irrigation) should be introduced to accelerate the evolution of reclaimed desert soil toward agricultural soil and to maintain the sustainability of the cropland ecosystem. Newly exploited wetland should be converted to meadow grassland, and the current extent of natural wetland should be protected from exploitation, to maintain its role of a carbon sink.

Acknowledgements

This study was financially supported by the National Natural Science Foundation of China (Project number: 91025018). Thanks go to staff of the State Key Laboratory of Soil Erosion and Dryland Farming on the Loess Plateau, Institute of Soil and Water Conservation, Chinese Academy of Sciences. We greatly appreciate editors of this journal and the three anonymous reviewers for their useful comments and suggestions.

References

Baritz, R., Seufert, G., Montanarella, L., Van Ranst, E., 2010. Carbon concentrations and stocks in forest soils of Europe. For. Ecol. Manag. 260, 262–277.
Batjes, N.H., 1996. Total carbon and nitrogen in the soils of the world. Eur. J. Soil Sci. 47, 151–163.
Batjes, N.H., 2006. Soil carbon stocks of Jordan and projected changes upon improved management of croplands. Geoderma 132, 361–371.
Batjes, N.H., Sombroek, W.G., 1997. Possibilities for carbon sequestration in tropical and subtropical soils. Glob. Change Biol. 3, 161–173.
Cambardella, C.A., Moorman, T.B., Novak, J.M., Parkin, T.B., Karlen, D.L., Turco, R.F., Konopka, A.E., 1994. Field-scale variability of soil properties in central Iowa soils. Soil Sci. Soc. Am. J. 58, 1501–1511.
De Vos, B., Van Meirvenne, M., Quataert, P., Deckers, J., Muys, B., 2005. Predictive quality of pedotransfer functions for estimating bulk density of forest soils. Soil Sci. Soc. Am. J. 69, 500–510.
Don, A., Schumacher, J., Scherer-Lorenzen, M., Scholten, T., Schulze, E.D., 2007. Spatial and vertical variation of soil carbon at two grassland sites—implications for measuring soil carbon stocks. Geoderma 141, 272–282.
Grace, J., 2004. Understanding and managing the global carbon cycle. J. Ecol. 92, 189–202.
Griffiths, R.P., Madritch, M.D., Swanson, A.K., 2009. The effects of topography on forest soil characteristics in the Oregon Cascade Mountains (USA): implications for the effects of climate change on soil properties. For. Ecol. Manag. 257, 1–7.
Grigal, D.F., Ohmann, L.F., 1992. Carbon storage in upland forests of the Lake States. Soil Sci. Soc. Am. J. 56, 935–943.
Grünzweig, J.M., Lin, T., Rotenberg, E., Schwartz, A., Yakir, D., 2003. Carbon sequestration in arid-land forest. Glob. Change Biol. 9, 791–799.
Han, F.P., Hu, W., Zheng, J.Y., Du, F., Zhang, X.C., 2010. Estimating soil organic carbon storage and distribution in a catchment of Loess Plateau, China. Geoderma 154, 261–266.
Harrison, A.F., Bocock, K.L., 1981. Estimation of soil bulk-density from loss-on-ignition values. J. Appl. Ecol. 18, 919–927.
Harriss, R.C., Sebacher, D.I., Day Jr., F.P., 1982. Methane flux in the Great Dismal Swamp. Nature 297, 673–674.
Hontoria, C., Rodríguez-Murillo, J.C., Saa, A., 1999. Relationships between soil organic carbon and site characteristics in Peninsular Spain. Soil Sci. Soc. Am. J. 614–621.
Jobbágy, E.G., Jackson, R.B., 2000. The vertical distribution of soil organic carbon and its relation to climate and vegetation. Ecol. Appl. 10, 423–436.
Karhu, K., Wall, A., Vanhala, P., Liski, J., Esala, M., Regina, K., 2011. Effects of afforestation and deforestation on boreal soil carbon stocks—comparison of measured C stocks with Yasso07 model results. Geoderma 164, 33–45.
Kaur, R., Kumar, S., Gurung, H.P., 2002. A pedo-transfer function (PTF) for estimating soil bulk density from basic soil data and its comparison with existing PTFs. Aust. J. Soil Res. 40, 847–857.
Lal, R., 2003. Soil erosion and the global carbon budget. Environ. Int. 29, 437–450.
Lal, R., 2008. Carbon sequestration. Philos. Trans. R. Soc. B Biol. Sci. 363, 815–830.
Leifeld, J., Bassin, S., Fuhrer, J., 2005. Carbon stocks in Swiss agricultural soils predicted by land-use, soil characteristics, and altitude. Agric. Ecosyst. Environ. 105, 255–266.
Li, X.G., Li, F.M., Rengel, Z., Bhupinderpal, S., Wang, Z.F., 2006. Cultivation effects on temporal changes of organic carbon and aggregate stability in desert soils of Hexi Corridor region in China. Soil Tillage Res. 91, 22–29.
Li, X.G., Li, Y.K., Li, F.M., Ma, Q.F., Zhang, P.L., Yin, P., 2009. Changes in soil organic carbon, nutrients and aggregation after conversion of native desert soil into irrigated arable land. Soil Tillage Res. 104, 263–269.
Limpens, J., Berendse, F., Blodau, C., Canadell, J.G., Freeman, C., Holden, J., Roulet, N., Rydin, H., Schaepman-Strub, G., 2008. Peatlands and the carbon cycle: from local processes to global implications — a synthesis. Biogeosci. Discuss. 1379–1419.
Lützow, M.v, Kögel-Knabner, I., Ekschmitt, K., Matzner, E., Guggenberger, G., Marschner, B., Flessa, H., 2006. Stabilization of organic matter in temperate soils: mechanisms and their relevance under different soil conditions — a review. Eur. J. Soil Sci. 57, 426–445.
Maia, S.M.F., Ogle, S.M., Cerri, C.C., Cerri, C.E.P., 2010. Changes in soil organic carbon storage under different agricultural management systems in the Southwest Amazon Region of Brazil. Soil Tillage Res. 106, 177–184.
Martin, D., Lal, T., Sachdev, C.B., Sharma, J.P., 2010. Soil organic carbon storage changes with climate change, landform and land use conditions in Garhwal hills of the Indian Himalayan mountains. Agric. Ecosyst. Environ. 138, 64–73.
Meersmans, J., De Ridder, F., Canters, F., De Baets, S., Van Molle, M., 2008. A multiple regression approach to assess the spatial distribution of Soil Organic Carbon (SOC) at the regional scale (Flanders, Belgium). Geoderma 143, 1–13.
Meersmans, J., van Wesemael, B., De Ridder, F., Fallas Dotti, M., De Baets, S., Van Molle, M., 2009a. Changes in organic carbon distribution with depth in agricultural soils in northern Belgium, 1960–2006. Glob. Change Biol. 15, 2739–2750.
Meersmans, J., van Wesemael, B., De Ridder, F., Van Molle, M., 2009b. Modelling the three-dimensional spatial distribution of soil organic carbon (SOC) at the regional scale (Flanders, Belgium). Geoderma 152, 43–52.
Mishra, U., Ussiri, D.A.N., Lal, R., 2010. Tillage effects on soil organic carbon storage and dynamics in Corn Belt of Ohio USA. Soil Tillage Res. 107, 88–96.
Moore, T.R., Dalva, M., 1993. The influence of temperature and water table position on carbon dioxide and methane emissions from laboratory columns of peatland soils. J. Soil Sci. 44, 651–664.
Norton, U., Mosier, A.R., Morgan, J.A., Derner, J.D., Ingram, L.J., Stahl, P.D., 2008. Moisture pulses, trace gas emissions and soil C and N in cheatgrass and native grass-dominated sagebrush-steppe in Wyoming, USA. Soil Biol. Biochem. 40, 1421–1431.
Post, W.M., Kwon, K.C., 2000. Soil carbon sequestration and land-use change: processes and potential. Glob. Change Biol. 6, 317–327.
Powers, J.S., Schlesinger, W.H., 2002. Relationships among soil carbon distributions and biophysical factors at nested spatial scales in rain forests of northeastern Costa Rica. Geoderma 109, 165–190.
Quinton, J.N., Govers, G., Van Oost, K., Bardgett, R.D., 2010. The impact of agricultural soil erosion on biogeochemical cycling. Nat. Geosci. 3, 311–314.
Rossi, J., Govaerts, A., De Vos, B., Verbist, B., Vervoort, A., Poesen, J., Muys, B., Deckers, J., 2009. Spatial structures of soil organic carbon in tropical forests—a case study of Southeastern Tanzania. Catena 77, 19–27.
Rustad, L.E., Fernandez, I.J., 1998. Experimental soil warming effects on CO_2 and CH_4 flux from a low elevation spruce–fir forest soil in Maine, USA. Glob. Change Biol. 4, 597–605.
Schimel, D.S., Parton, W.J., 1986. Microclimatic controls of nitrogen mineralization and nitrification in shortgrass steppe soils. Plant Soil 93, 347–357.
Sleutel, S., De Neve, S., Hofman, G., 2003. Estimates of carbon stock changes in Belgian cropland. Soil Use Manag. 19, 166–171.
Su, Y.Z., Zhao, W.Z., Su, P.X., Zhang, Z.H., Wang, T., Ram, R., 2007. Ecological effects of desertification control and desertified land reclamation in an oasis–desert ecotone in an arid region: a case study in Hexi Corridor, northwest China. Ecol. Eng. 29, 117–124.
Su, Y.Z., Liu, W.J., Yang, R., Chang, X.X., 2009. Changes in soil aggregate, carbon, and nitrogen storages following the conversion of cropland to alfalfa forage land in the marginal oasis of northwest China. Environ. Manag. 43, 1061–1070.
Su, Y.Z., Yang, R., Liu, W.J., Wang, X.F., 2010. Evolution of soil structure and fertility after conversion of native sandy desert soil to irrigated cropland in arid region, China. Soil Sci. 175, 246–254.
Su, Y.Z., Wang, X.F., Yang, R., Yang, X., Liu, W.J., 2012. Soil fertility, salinity and nematode diversity influenced by *Tamarix ramosissima* in different habitats in an arid desert oasis. Environ. Manag. 50, 226–236.
Tan, Z.X., Lal, R., Smeck, N.E., Calhoun, F.G., 2004. Relationships between surface soil organic carbon pool and site variables. Geoderma 121, 187–195.
Thomas, K.L., Benstead, J., Davies, K.L., Lloyd, D., 1996. Role of wetland plants in the diurnal control of CH_4 and CO_2 fluxes in peat. Soil Biol. Biochem. 28, 17–23.
Trumbore, S., 2000. Age of soil organic matter and soil respiration: radiocarbon constraints on belowground C dynamics. Ecol. Appl. 10, 399–411.
Walkley, A., Black, I.A., 1934. An examination of the Degtjareff method for determining organic carbon in soils: effect of variations in digestion conditions and of inorganic soil constituents. Soil Sci. 63, 251–263.
Wang, Y.Q., Shao, M.A., Liu, Z.P., 2010. Large-scale spatial variability of dried soil layers and related factors across the entire Loess Plateau of China. Geoderma 159, 99–108.
Wang, Y.F., Fu, B.J., Lü, Y.H., Chen, L.D., 2011. Effects of vegetation restoration on soil organic carbon sequestration at multiple scales in semi-arid Loess Plateau, China. Catena 85, 58–66.
Wang, S.F., Wang, X.K., Ouyang, Z.Y., 2012a. Effects of land use, climate, topography and soil properties on regional soil organic carbon and total nitrogen in the Upstream Watershed of Miyun Reservoir, North China. J. Environ. Sci. 24, 387–395.
Wang, Z., Liu, G.B., Xu, M.X., Zhang, J., Wang, Y., Tang, L., 2012b. Temporal and spatial variations in soil organic carbon sequestration following revegetation in the hilly Loess Plateau, China. Catena 99, 26–33.
Wei, X.R., Qiu, L.P., Shao, M.A., Zhang, X.C., Gale, W.J., 2012. The accumulation of organic carbon in mineral soils by afforestation of abandoned farmland. PLoS ONE 7, e32054.
West, N.E., Stark, J.M., Johnson, D.W., Abrams, M.M., Wight, J.R., Heggem, D., Peck, S., 1994. Effects of climatic change on the edaphic features of arid and semiarid lands of western North America. Arid Soil Res. Rehabil. 8, 307–351.
Yang, X.Y., Ren, W.D., Sun, B.H., Zhang, S.L., 2012. Effects of contrasting soil management regimes on total and labile soil organic carbon fractions in a loess soil in China. Geoderma 177–178, 49–56.
Zhang, J.H., Li, G.D., Nan, Z.R., Xiao, H.L., 2012. Research on soil particle size distribution and its relationship with soil organic carbon under the effects of tillage in the Heihe oasis. Geogr. Res. 31, 608–618 (in Chinese with English abstract).
Zhao, Y., Chai, Q., Chen, Y.Y., Zhu, W.H., 2008. Improvement and utilization of saline-alkali grassland in Hexi Corridor. Pratacult. Sci. 25, 21–25 (in Chinese with English abstract).
Zinn, Y.L., Lal, R., Resck, D.V.S., 2005. Texture and organic carbon relations described by a profile pedotransfer function for Brazilian Cerrado soils. Geoderma 127, 168–173.