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# Profile Distributions and Controls of Soil Inorganic Carbon along a 150-Year Natural Vegetation Restoration Chronosequence

**Soil inorganic C (SIC) comprises approximately a third of the global soil C pool, which plays an important role in global C cycling. However, there is still considerable disagreement on the direction and magnitude of changes in SIC stocks following vegetation restoration. We conducted a study comparing SIC at different succession stages along a 150-yr natural vegetation restoration chronosequence to examine the effect of long-term natural vegetation restoration on the distribution of SIC and to identify the factors that control changes in SIC. The results showed that SIC storage in the top 10 cm gradually decreased (0.092 Mg ha$^{-1}$ yr$^{-1}$) along the vegetation restoration chronosequence but was basically unchanged in the subsoil (10–100 cm). The soil total C storages remained stable across the restoration sequence, with the redistribution of the total soil C pool from SIC to soil organic C (SOC) the dominant forms. The controlling factors of SIC were different in the top- and subsoils along the chronosequence. In the top 30-cm soil layers, SOC was a good predictor of SIC; however, for soils below 30 cm, soil sand and silt contents and pH were better predictors of SIC. Therefore, we can conclude that the variations in SOC induced by vegetation restoration were the main driving force for SIC changes in the topsoil, while the genetic soil features (i.e., sand and silt contents) were the controlling factors that determined the amount of SIC in the subsoil.**

Abbreviations: BD, bulk density; PIC, pedogenic inorganic carbon; RD, root mass density; SIC, soil inorganic carbon; SICD, soil inorganic carbon storage; SOC, soil organic carbon; SOCD, soil organic carbon storage; SWC, soil water content; STCD, soil total carbon storage.

Soil is the largest C pool (2500 Pg) in the terrestrial ecosystem and is approximately twice the size of the sum of the biotic (560 Pg) and atmospheric (760 Pg) C pools (Lal, 2004). Even a minor change in the soil C pool would have a great impact on the atmospheric $CO_2$ concentration and would influence global warming (Trumbore and Czimczik, 2008; Yang et al., 2010). Accurately estimating C storage and depth to better understand the process of C cycling in soils in various ecosystems is critical for modeling and regulating the terrestrial C cycle (Zhang et al., 2015).

The soil C pool is comprised of the SOC and SIC pools. It has been reported that the global SOC pool ranges from 1200 to 1600 Pg (Batjes, 1996; Eswaran et al., 1993), and the estimated global SIC pool varies from a range of 695 to 748 Pg (Batjes, 1996) to 1738 Pg (Eswaran et al., 1995) in the top 1-m layer. Because of its potentially rapid response to climate and land use changes, and

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## Core Ideas:

- **Topsoil SIC storage decreased with restoration age, but subsoils were unchanged.**
- **SOC was the main driving force of SIC storage in the topsoil.**
- **Sand and silt contents were the controlling factors of SIC storage in subsoils.**
- **Soil total C storage remained stable across the restoration sequence.**

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thus its influence on global warming, the SOC pool has received considerable attention and has been extensively researched worldwide in recent decades (Liu et al., 2014; McBratney et al., 2014). In contrast with the great progress made in understanding the SOC pool, the SIC pool has been largely neglected in global C budgets, mainly because of the small amount of SIC exchange with the atmosphere but also due to its restricted distribution around the world (Monger, 2014; Wang et al., 2010b). However, SIC comprises approximately a third of the global soil C pool and plays an important role in global C cycling (Lal, 2004). Some recent studies have reported that the accumulation rate of SIC in some regions is much higher than previously believed. Ryskov et al. (2008) reported SIC accumulation at a rate of 0.86 to 2.2 kg C $m^{-2}$ $yr^{-1}$ in soils in Russia during the last 5000 yr. Likewise, SIC absorption by saline and alkaline soils in western China could be as high as 62 to 622 g C $m^{-2}$ $yr^{-1}$, which is one or two orders of magnitude higher than the previously reported value of 1 to 5 g C $m^{-2}$ $yr^{-1}$ (Xie et al., 2009). However, these high accumulation rates of SIC have been questioned by some researchers who have not observed such high values (Liu, 2011; Schlesinger et al., 2009). Therefore, there is an urgent need to further understand the role of SIC in global C cycling and also its impact on climate change.

The SIC pool can be divided into lithogenic inorganic C and pedogenic inorganic C (PIC). The former is inherited from the soil parent material, and the latter is formed through the dissolution and precipitation of secondary carbonate material (Liu et al., 2014; Wu et al., 2009). The formation of PIC can be expressed in the following chemical formulas:

$$CaCO_3 + H_2O + CO_{2\downarrow} \rightarrow Ca^{2+} + 2HCO_3^- \qquad [1]$$

$$Ca^{2+} + 2HCO_3^- \rightarrow CaCO_3 + H_2O + CO_{2\uparrow} \qquad [2]$$

$$CaSiO_3 + 3H_2O + 2CO_{2\downarrow} \rightarrow Ca^{2+} + 2HCO_3^- + H_4SiO_4 \qquad [3]$$

$$Ca^{2+} + 2HCO_3^- \rightarrow CaCO_3 + H_2O + CO_{2\uparrow} \qquad [4]$$

Pedogenic inorganic C is formed during the dissolution and reprecipitation of $CaCO_3$ or $CaSiO_3$ parent material (Liu et al., 2014; Wu et al., 2009). According to Eq. [1] and [2], 1 mol of atmospheric $CO_2$ is consumed during $CaCO_3$ dissolution, but an equal amount is liberated during pedogenic carbonate precipitation. Pedogenic inorganic C originating from the weathering of $CaSiO_3$ could consume 2 mol of atmospheric $CO_2$ for every 1 mol released during pedogenic carbonate precipitation (Eq. [3] and [4]), which results in a net increase in SIC (Monger and Martinez-Rios, 2000; Wu et al., 2009). The dissolved inorganic C transported to the stream system through surface and subsurface runoff or the precipitation of external $Ca^{2+}$ imported from rain and dust can also produce a SIC sink (Chang et al., 2012; Jin et al., 2014; Mikhailova et al., 2013).

Land use changes are the next largest source of atmospheric $CO_2$ after the consumption of fossil fuels (Lal, 2004). From 1959 to 2006, the average $CO_2$ emission from land use changes was about 1.5 ± 0.5 Pg C $yr^{-1}$, which contributed 22.4% of the total emissions from human activities (Canadell et al., 2007). Many previous studies have addressed the strong influence of land use change on the storage and dynamics of SOC (Chen et al., 2007; Deng et al., 2014a; Guo and Gifford, 2002; Hernandez-Ramirez et al., 2011; Poeplau and Don, 2013) and have indicated that SOC storage increases after cropland is converted to forest or grassland (Deng et al., 2014a), while it significantly declines through the conversion of native forest to cropland or plantation (Don et al., 2011; Guo and Gifford, 2002; Poeplau and Don, 2013). Some studies have indicated that land use changes also have a great impact on the SIC pool (Jelinski and Kucharik, 2009; Wu et al., 2009). Great differences in SIC storage were observed under different types of land use (Liu et al., 2014; Mi et al., 2008; Wang et al., 2015). Land use changes could lead to the redistribution of SIC due to the differences in $CO_2$ concentration, soil water content, and pH in soil profiles under different types of land use (Chang et al., 2012; Liu et al., 2014). However, to the best of our knowledge, the impact of land use changes on SIC storage and dynamics have not been studied as extensively as their impact on SOC storage (Liu et al., 2014; Mi et al., 2008). Therefore, SIC dynamics and the processes involved in its accumulation or loss from soils after vegetation restoration needs to be better understood.

The Loess Plateau has the highest SIC content in China (Wu et al., 2009). The mean SIC density in the 0- to 100-cm soil layer on the Loess Plateau is 17.04 kg C $m^{-2}$, which is about three times that of the national average (6.3 kg C $m^{-2}$), and the corresponding SIC pool is approximately 10.2 Pg C, which represents approximately 18.4% of the total SIC pool in China (55.3 Pg C) (Tan et al., 2014; Wu et al., 2009). The Loess Plateau is also well known for its serious erosion problems, with the mean soil erosion rate ranging from 5000 to 10000 Mg $km^{-2}$ $yr^{-1}$ (Wang et al., 2013a). To control soil erosion, large-scale vegetation restoration projects have been launched in the last few decades, which have caused great changes in land use on the Loess Plateau (Deng et al., 2014a). The effects of vegetation restoration on SOC sequestration have been extensively investigated at the local and regional scales (Chang et al., 2011; Chen et al., 2007; Deng et al., 2013; Wang et al., 2011a). In contrast, only a few studies have focused on the variations and controlling factors of the SIC pool after vegetation restoration on the Loess Plateau (Chang et al., 2012; Liu et al., 2014). Therefore, the objectives of this study were to: (i) examine the effect of long-term natural vegetation restoration on the soil profile characteristics of SIC; and (ii) identify the factors that control the changes in SIC.

## MATERIALS AND METHODS

### Study Area

The study was conducted at Lianjiabian Forest Farm, which is located in the Ziwuling Mountain region in the center of the Loess Plateau (Fig. 1). The study area has landforms typi-

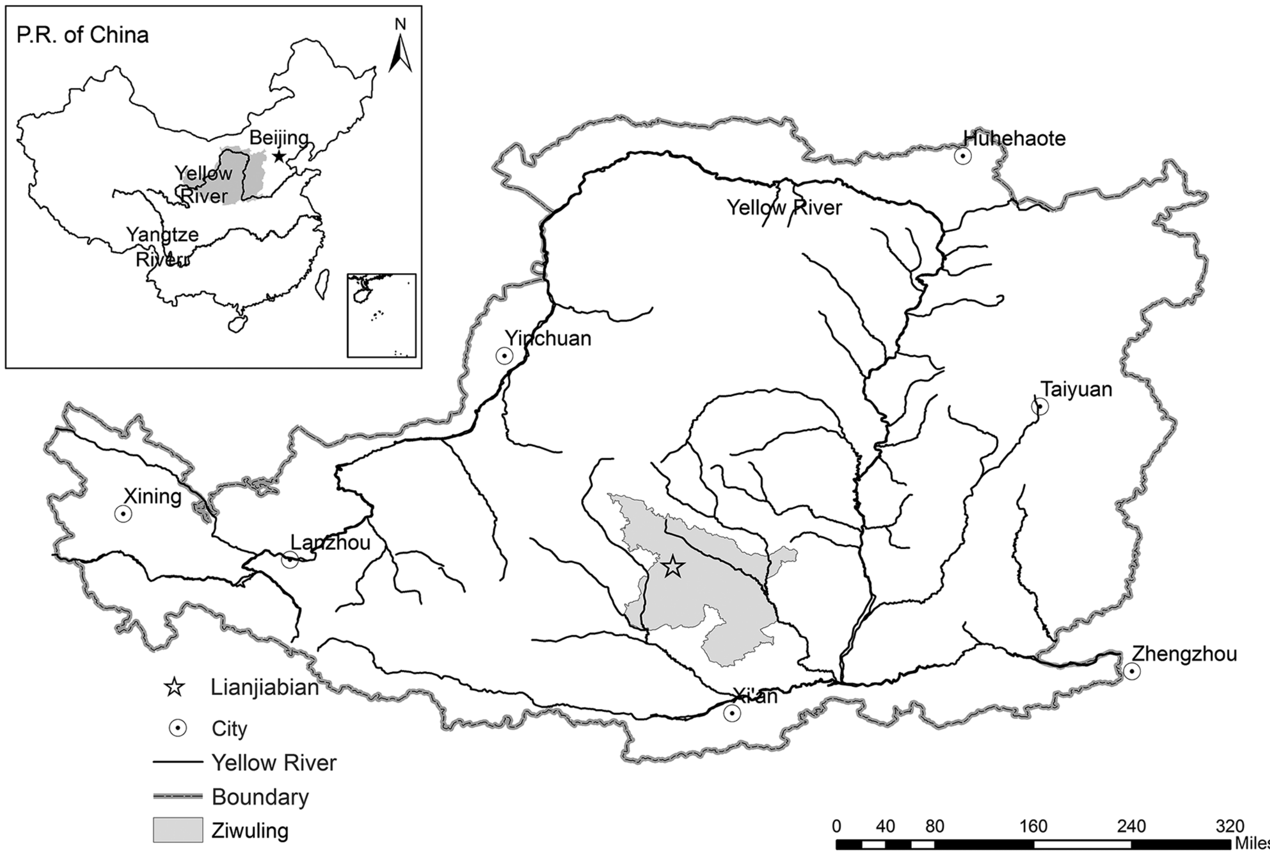


**Fig. 1. Location of the study sites on the Loess Plateau. The shaded area in the upper left corner of the map represents the range of the Loess Plateau in China, and the shaded area in the main map represents the range of the Ziwuling Mountains within the Loess Plateau.**

cal of loess hilly topography and a mid-temperate continental monsoon climate. The annual mean temperature is 10°C and the mean annual precipitation is 587 mm (1960–2010) (Deng et al., 2014b). The soils are largely loessial and have developed from primary or secondary loess parent materials, which are evenly distributed 50 to 130 m deep above a red earth classified as a Haplustalf, with soil pH ranges from 7.92 to 8.31 (Deng et al., 2014b). The natural biomes are deciduous broadleaf forests, in which the climax vegetation is *Quercus liaotungensis* Koitz. (Liaotung oak) forest; *Populus davidiana* Dode (Korean aspen) and *Betula platyphylla* Sukaczev (Asian white birch) communities are distributed as the pioneer forests; *Sophora davidii* (Franch.) Skeels (David's mountain laurel), *Hippophae rhamnoides* L. (sea-buckthorn), *Rosa xanthina* Lindl. (yellow rose), and *Spiraea pubescens* Turcz. (meadow sweets) are the main shrub species; *Bothriochloa ischaemun* (L.) Keng (yellow bluestem), *Carex lanceolata* Boott (sedge), *Potentilla chinensis* Seringe (Chinese cinquefoil), and *Stipa bungeana* Trin. (needlegrass) are the main herbaceous species. The main crop types in this region are *Setaria italica* (L.) P. Beauv. (foxtail millet), *Zea mays* L. (corn), *Solanum tuberosum* L. (potato), and *Glycine max* (L.) Merr. (soybean) (Wang et al., 2010a).

## Site Selection

The Ziwuling Mountains region contains a complete sequence of natural vegetation succession. Such a sequence is rarely found on the Loess Plateau. Natural vegetation at different stages in the restoration process is found in this region (Wang et al., 2010a). The ages of the communities were determined by two methods. For shrub and herbaceous communities that regenerated for no more than 60 yr, we verified the restoration age by interviewing local elders. For the forest communities with >60 yr recovery, we determined the restoration age by referring to related written sources (Zou et al., 2002). The more detailed description of the method used to identify the ages of the communities can be found in previous studies (Wang et al., 2010a; Deng et al., 2013, 2014b; Zhao et al., 2015).We selected five communities at different stages of restoration (11, 35, 60, 100, and 150 yr) to evaluate the effect of long-term natural vegetation restoration on SIC storage on the Loess Plateau. For comparison, one farmland plot planted with corn (*Zea mays* L.) was selected as a baseline or control (0 yr). To minimize the effects of site conditions on the experimental results, all sites selected had a similar slope aspect, gradient, elevation, soil type, and land use history. The details for each site are listed in Table 1.

## Soil Sampling and Analysis

Soil sampling was conducted in November 2013. Three plots were randomly chosen at each stage of restoration. The plots were 20 by 20 m in the forest community, 5 by 5 m in the shrub communities, and 2 by 2 m in the herbaceous communities. The litter in each plot was cleared before sampling. Soil

**Table 1. Restoration stage, geographical location, topography, and vegetation for the sampling sites. R0 is farmland; R11, R35, R60, R100, and R150 are vegetation communities at different stages of restoration.**

| Restoration stage | Age | Longitude | Latitude | Elevation | Slope | Coverage† | Dominant community | Land use history |
|---|---|---|---|---|---|---|---|---|
| | yr | | | m | ° | % | | |
| R0 | 0 | 108°28′5.36″ E | 36°4′24.87″ N | 1476 | 10–15 | 75 | *Zea mays* L. | farmland |
| R11 | 11 | 108°31′57.86″ E | 36°5′3.65″ N | 1352 | 10–15 | 40 | *Acanthopanax giraldii* Harms | farmland |
| R35 | 35 | 108°31′38.52″ E | 36°5′8.83″ N | 1346 | 8–12 | 85 | *Bothriochloa ischaemun* (L.) Keng | farmland |
| R60 | 60 | 108°31′53.1″ E | 36°4′57.99″ N | 1334 | 10–15 | 85 | *Hippophae rhamnoides* L. | farmland |
| R100 | 100 | 108°31′44.78″ E | 36°2′54.46″ N | 1449 | 10–15 | 85 | *Populus davidiana* Dode | farmland |
| R150 | 150 | 108°32′13.57″ E | 36°2′56.01″ N | 1437 | 10–15 | 90 | *Quercus liaotungensis* Koitz. | farmland |

† The coverage of each community is the maximum value in the vegetation growth period. The land use history of each site was determined by consulting the literature, county annals, and local elders.

samples were collected with a 9-cm i.d. corer at the four corners and center of each plot. The mineral soil layer was collected at intervals of 0 to 10, 10 to 20, 20 to 30, 30 to 50, 50 to 70, and 70 to 100 cm, and the samples were mixed to produce one sample for each layer. All the soil samples were air dried and sieved through a 2-mm screen, and roots and other debris were removed. The bulk density (BD) of each soil layer was measured using a soil bulk sampler with a 5-cm-diameter and 5-cm-high stainless steel cutting ring, with three replicates in each plot. Soil water content (SWC) was measured gravimetrically and expressed as a percentage of soil water to dry soil weight. Soil pH was determined using a soil/water ratio of 1:2.5 (PHSJ-4A pH meter, Zhangqiu Meihua International Trading Co.). The grain size of soil samples was determined by laser granulometry (Mastersizer 2000, Malvern Instruments Ltd.). The roots at each depth from 0 to 100 cm were also sampled and oven dried to measure the root mass density (RD). The SIC was analyzed using an inorganic C analyzer (CM140, UIC Inc.). The SOC content was measured using the dichromate oxidation method (Kalembasa and Jenkinson, 1973). The basic soil properties are shown in Table 2.

Because there was no coarse fraction (>2 mm) in the soil samples, the SIC and SOC storage (Mg ha$^{-1}$) was calculated using (Jin et al., 2014)

$$\mathrm{SICD}_i = \sum_{i=1}^{n} \frac{D_i \times \mathrm{BD}_i \times \mathrm{SIC}_i}{10} \qquad [5]$$

$$\mathrm{SOCD}_i = \sum_{i=1}^{n} \frac{D_i \times \mathrm{BD}_i \times \mathrm{SOC}_i}{10} \qquad [6]$$

where $\mathrm{SICD}_i$ and $\mathrm{SOCD}_i$ are the SIC and SOC storage of the $i$th layer; $D_i$ (cm), $\mathrm{BD}_i$ (g cm$^{-3}$), $\mathrm{SIC}_i$ (g kg$^{-1}$), and $\mathrm{SOC}_i$ (g kg$^{-1}$) are the soil thickness, bulk density, SIC content and SOC content of the $i$th layer, respectively; and 10 is the unit conversion factor. The soil total C storage (STCD) (Mg ha$^{-1}$) was calculated as the sum of the SIC and SOC storage.

## Statistical Analysis

One-way ANOVA was used to analyze the differences in the SICD among the different stages of restoration. Differences were evaluated at the 5% significance level. When a testing for homogeneity of variance was passed and significance was observed at the $P < 0.05$ level, a least significant difference test was used for multiple comparisons. A generalized linear model was adopted to determine whether there were significant correlations between SIC, SICD, SOCD, STCD, and restoration ages. Pearson's test was used to examine the correlations between SIC, SICD, and soil and vegetation properties for the data from all stages of vegetation restoration. Stepwise regression was used to analyze the dominant factors that impact on the SIC and SICD dynamics across the vegetation restoration chronosequence. All statistical analyses were performed using the software program SPSS, Version 16.0 (SPSS Inc.).

# RESULTS

## Profile Distribution of Soil Inorganic Carbon Storage

The SICD significantly decreased at a soil depth of 0 to 10 cm, slightly fluctuated in the soil layers at 10 to 50 cm, and was basically unchanged below the 50-cm soil depth across the 150-yr vegetation restoration chronosequence (Fig. 2). The SICD declined from 20.4 to 7.2 Mg ha$^{-1}$ in the 0- to 10-cm soil layer, while it changed from 71.8 to 69.4 Mg ha$^{-1}$ in the 70- to 100-cm soil layer across the chronosequence. The SICD at depths

**Table 2. Basic soil information at the 0- to 10-cm depth for the sampling sites.**

| Restoration stage | Soil organic C | Bulk density | Soil water content | pH | Clay | Silt | Sand |
|---|---|---|---|---|---|---|---|
| | g kg$^{-1}$ | g cm$^{-3}$ | % | | | % | |
| R0 | 10.75 | 1.41 | 9.89 | 7.69 | 6.41 | 76.45 | 17.14 |
| R11 | 11.62 | 1.38 | 12.85 | 7.50 | 6.19 | 80.59 | 13.21 |
| R35 | 15.93 | 1.21 | 13.34 | 7.46 | 6.22 | 79.69 | 14.09 |
| R60 | 20.72 | 1.22 | 12.67 | 7.26 | 7.57 | 78.85 | 13.59 |
| R100 | 26.30 | 1.01 | 16.17 | 7.27 | 5.62 | 75.95 | 18.42 |
| R150 | 35.15 | 1.02 | 14.46 | 7.53 | 5.92 | 77.33 | 16.75 |


**Fig. 2. Soil inorganic C (SIC) storage in each layer at different stages of restoration. Different lowercase letters denote significant differences in each layer's SIC storage among the different stages of restoration; different uppercase letters denote significant differences in SIC storage of the 0- to 100-cm depth among the different stages of restoration (*P* < 0.05). Error bars are the standard error (*n* = 3).**

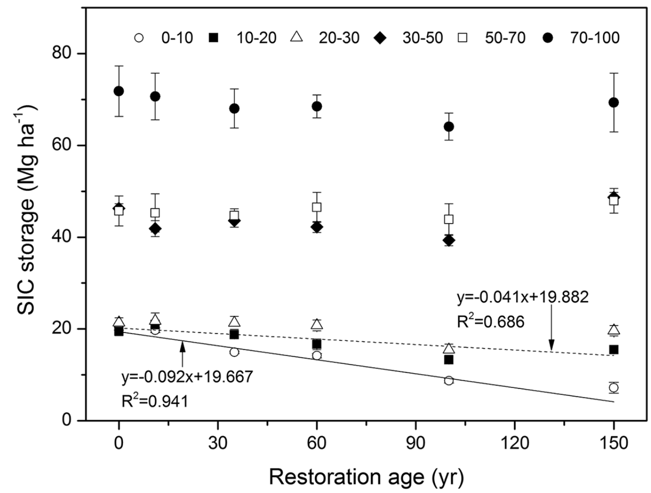


**Fig. 3. Changes in soil inorganic C (SIC) storage in each layer along a 150-yr natural vegetation restoration chronosequence. Linear fitting was adopted when the regression equation between SIC storage and restoration age was significant (*P* < 0.05).**

of 0 to 100 cm also exhibited a significant decrease, with a decline from 225.1 Mg ha$^{-1}$ in farmland (0 yr) to 208.3 Mg ha$^{-1}$ under *Quercus liaotungensis* forest (150 yr). The SICD in the 0- to 10- and 10- to 20-cm layers had a significantly negative relationship with the duration of the restoration, with the rate of decrease in SIC storage being 0.092 and 0.041 Mg ha$^{-1}$ yr$^{-1}$, respectively (Fig. 3).

## Contribution of Soil Inorganic Carbon Storage to Soil Total Carbon Storage

The total soil C storage did not show a significant increase at either 0 to 10 or 0 to 100 cm across the vegetation restoration chronosequence (Fig. 4). The STCD ranged from 34.2 to 43.3 and 275.0 to 308.8 Mg ha$^{-1}$ at soil depths of 0 to 10 and 0 to 100 cm, respectively (Fig. 4). The contribution of SICD to the soil C pool clearly increased with an increase in soil depth. It ranged from 38.9% at 0 to 10 cm to 71.5% at 0 to 100 cm (Fig. 5). However, the contribution of SIC storage to the soil C pool gradually decreased across the chronosequence, with the mean value changing from 65.2% in farmland (0 yr) to 43.3% under a *Q. liaotungensis* forest (150 yr) (Fig. 5).

## Relationship of Soil Inorganic Carbon Content and Storage with Soil and Vegetation Properties

Both SICD and SIC had different relationships with soil and vegetation properties in the different soil layers (Table 3). In the 0- to 10-, 10- to 20-, and 20- to 30-cm soil layers, SICD was positively correlated with SIC, BD, and pH but negatively correlated with SOC and RD (*P* < 0.05). In addition, SICD was also negatively correlated with SWC in the 0- to 10-cm layer and with the sand content in the 20- to 30-cm layer but positively correlated with the SWC and silt content in the 20- to 30-cm layer (*P* < 0.05). However, SICD only had a significant positive correlation with BD in the 30- to 50- and 70- to 100-cm soil

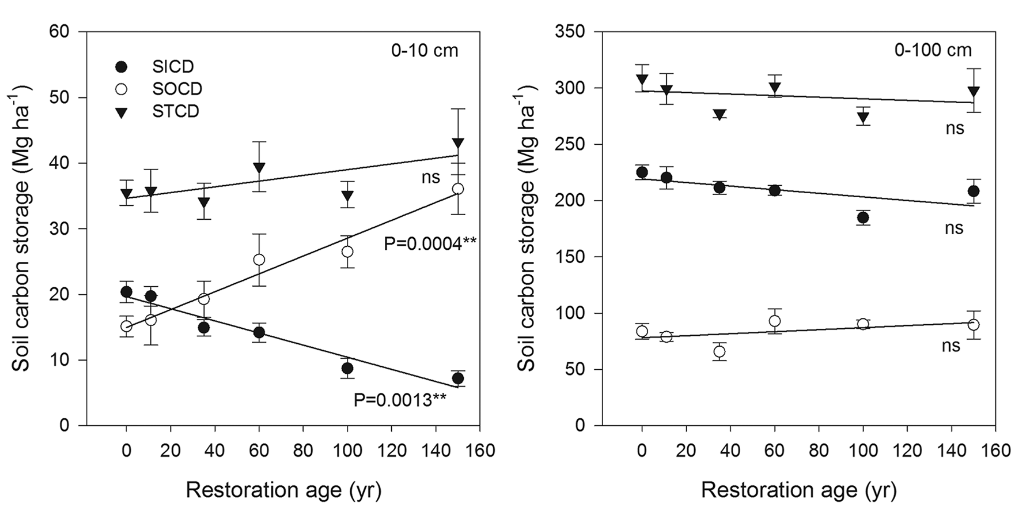


**Fig. 4. Changes in soil inorganic (SICD), organic (SOCD), and total (STCD) C storage at the 0- to 10- and 0- to 100-cm soil depths across a 150-yr natural vegetation restoration chronosequence. Error bars are the standard error (n = 3). ***P* < 0.01; ns, not significant (*P* > 0.05).**

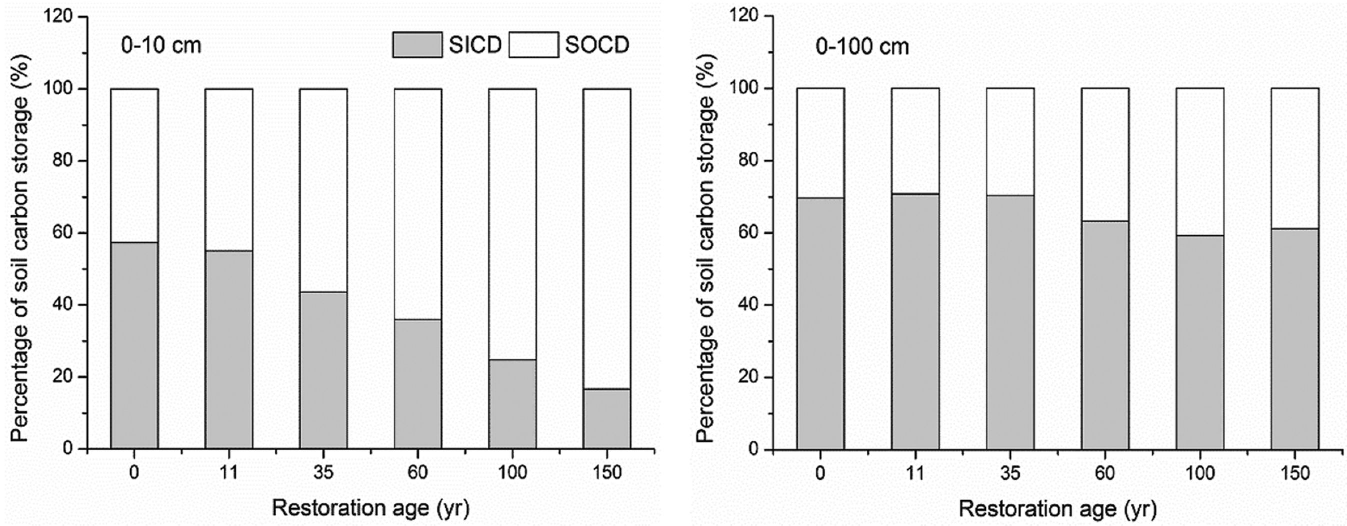


**Fig. 5. Percentages of soil inorganic (SICD) and organic (SOCD) C storage at the 0- to 10- and 0- to 100-cm soil depths across a 150-yr natural vegetation restoration chronosequence.**

layers ($P < 0.01$), while it was positively correlated with SIC and BD in the 50- to 70-cm soil layer ($P < 0.05$).

The relationships of SIC with the soil and vegetation properties was similar to the relationships with SICD (Table 3). The SIC was positively correlated with BD and negatively correlated with SOC and RD in the 0- to 10- and 10- to 20-cm layers ($P < 0.05$). In soil layers below 20 cm, SIC was positively correlated with the silt content and negatively correlated with the sand content in the 20- to 30-, 30- to 50-, and 50- to 70-cm soil layers. In addition, SIC was also positively correlated with pH in the 10- to 20-, 20- to 30-, and 70- to 100-cm soil layers but negatively correlated with SOC in the 20- to 30- and 30- to 50-cm layers ($P < 0.05$).

The factors that affected SIC and SICD were further screened using a stepwise regression method. Soil organic C had good predictive power for SIC in the 0- to 10-, 10- to 20-, and 20- to 30-cm layers; however, soil properties such as the sand and silt contents and pH were good predictors of SIC in the layers below 30 cm (Table 4). For SICD, SOC and SWC were good predictors in the 0- to 10-cm layer, while RD and sand content were good predictors in the 10- to 20- and 20- to 30-cm layers, respectively (Table 4). However, no variables entered into the stepwise regression equations at soil depths of 30 to 50, 50 to 70, and 70 to 100 cm, which indicated that no dominant factors could determine the changes in SICD in soil layers below 30 cm.

## DISCUSSION

### Effect of Vegetation Restoration on Soil Inorganic Carbon Content and Storage

The SIC and SICD in the 0- to 100-cm layer in the study area ranged from 16.2 to 14.0 g kg$^{-1}$ and 225.1 to 184.8 Mg ha$^{-1}$, respectively (Fig. 3 and 6). Both SIC and SICD in the 0- to 10-cm layer showed a significant decline across the vegetation restoration chronosequence, with a rate of decrease of 0.052

**Table 3. Correlations between the soil inorganic C content (SIC) and storage (SICD) and the soil properties of organic C content (SOC), bulk density (BD), soil water content (SWC), pH, particle size distribution (clay, silt, and sand contents), and root mass density (RD). Three replications were used at each site to conduct the correlation analysis. Root mass density in the farmland was not measured because the corn was harvested during the sampling period; therefore, the sample size for RD was 15 while for the others it was 18.**

| Soil layer | Parameter | SIC | SOC | BD | SWC | pH | Clay | Silt | Sand | RD |
|---|---|---|---|---|---|---|---|---|---|---|
| cm | | | | | | | | | | |
| 0–10 | SICD | 0.989** | −0.940** | 0.982** | −0.656** | 0.474* | 0.304 | 0.314 | −0.374 | −0.729** |
| | SIC | 1 | −0.965** | 0.951** | −0.604** | 0.389 | 0.313 | 0.345 | −0.405 | −0.750** |
| 10–20 | SICD | 0.926** | −0.695** | 0.878** | −0.213 | 0.766** | −0.231 | 0.480* | −0.357 | −0.718** |
| | SIC | 1 | −0.688** | 0.636** | −0.294 | 0.668** | −0.181 | 0.438 | −0.336 | −0.605* |
| 20–30 | SICD | 0.841** | −0.539* | 0.805** | 0.655** | 0.693** | 0.063 | 0.737** | −0.772** | −0.558* |
| | SIC | 1 | −0.703** | 0.358 | 0.466 | 0.521* | −0.015 | 0.632** | −0.640** | −0.476 |
| 30–50 | SICD | 0.362 | −0.264 | 0.884** | 0.093 | 0.159 | 0.166 | 0.066 | 0.109 | 0.210 |
| | SIC | 1 | −0.489* | −0.113 | 0.124 | 0.431 | −0.182 | 0.671** | −0.645** | −0.259 |
| 50–70 | SICD | 0.558* | −0.105 | 0.908** | 0.220 | 0.090 | 0.090 | 0.268 | −0.298 | 0.041 |
| | SIC | 1 | −0.168 | 0.161 | 0.195 | 0.443 | 0.117 | 0.741** | −0.784** | −0.334 |
| 70–100 | SICD | 0.438 | 0.059 | 0.872** | −0.131 | 0.269 | −0.344 | 0.365 | −0.261 | −0.077 |
| | SIC | 1 | 0.232 | −0.056 | −0.541 | 0.617** | −0.392 | 0.550* | −0.442 | −0.621 |

* Correlation is significant at the 0.05 level.

** Correlation is significant at the 0.01 level (two-tailed).

g $kg^{-1}$ $yr^{-1}$ and 0.092 Mg $ha^{-1}$ $yr^{-1}$, respectively. However, no clear trends in SIC and SICD with the duration of restoration were observed below a soil depth of 10 cm (Fig. 3 and 6). This result agrees with previous studies that have indicated that vegetation restoration has a significant effect on SIC redistribution throughout the soil profile. For example, Chang et al. (2012) found that the SICD in the top 20 cm of a forest soil was significantly lower than that in the equivalent layer of a farmland soil. Liu et al. (2014) also reported a significant decrease in both the SIC and SICD at depths of 0 to 80 cm after farmland was converted to grassland. In contrast, the SIC and SICD of farmland in the 0- to 25-cm soil layer were significantly higher than in a prairie soil on a floodplain in the midwestern United States (Jelinski and Kucharik, 2009). However, differences have been reported among studies in the soil depth where SIC significantly declines. In the Loess Plateau, Chang et al. (2012) reported that the SIC decreased significantly in the top 20 cm, which was in line with our study but was much shallower than the depth reported by Liu et al. (2014). The reasons for these differences might be due to differences in the SOC and RD. Wei et al. (2012) reported that the SOC in the 0- to 100-cm layer of restored grasslands was more than twice that of restored forests, and the SOC in each soil layer below 10 cm in grassland was also higher than that in the forest. A large SOC pool often results in a looser soil structure and higher soil permeability, which aids the dissolution and leaching of carbonate throughout the profile.

Vegetation restoration could also affect the storage of SIC. However, the exact extent of the impact of revegetation on the SICD in the soil profile could not be determined. Liu et al. (2014) found a significant decrease in SICD in a 0- to 200-cm profile after farmland was restored to grassland and the trend of a declining SICD was enhanced in the later stages of restoration. Similarly, a decreasing pattern of SICD was also observed with an increase in plantation age (Sartori et al., 2007; Wang et al., 2013b). However, Chang et al. (2012) reported that the decrease in SICD in the top 20 cm of a forest soil was counterbalanced by an increase in SICD below 20 cm. Thus, there was no SICD sequestration in the 0- to 100-cm profile after the conversion of farmland to forest. These inconsistent results from previous studies indicate that there must be multiple factors working together to determine the changes in SICD in the soil profile. The SICD in the 0- to 100-cm profile in our study generally declined with age of restoration, although there was a higher SICD in the *Q. liaotungensis* forest (150 yr) than in the *P. davidiana* forest (100 yr) (Fig. 2 and 5). Previous studies have shown that the carbonates in soil could be lost through three potential pathways: (i) emission into the atmosphere as $CO_2$; (ii) dissolution and leaching into the subsoil or groundwater; and (iii) translocation to nearby regions with surface runoff (Liu et al., 2014; Yang et al., 2012). Although the SICD in the top 10 cm of restored vegetation was significantly decreased compared with the farmland, no increase in SICD was observed in the subsoil under the restored vegetation (Fig. 2). This indicates that the dissolution and leaching of SIC made little contribution to the SICD loss under restored vegetation. Moreover, during vegetation restoration, a mass of soil $Ca^{2+}$ accumulated in plant tissues, which led to a large decline in soil $Ca^{2+}$ (McLaughlin and Wimmer, 1999). The decrease in soil $Ca^{2+}$ would inevitably cause the transforma-

**Table 4. Stepwise regression of soil inorganic C content (SIC) and storage (SICD) with soil and vegetation properties in different soil layers across a 150-yr vegetation restoration chronosequence. The regression relationship was significant at *P* < 0.05. No independent variables entered into the stepwise regression equations with the SICD as the dependent variable at soil depths of 30 to 50, 50 to 70, and 70 to 100 cm.**

| Soil layer | Equation† | $R^2$ | Sign. | *n* |
|---|---|---|---|---|
| cm | | | | |
| 0–10 | SIC = 4.61 − 0.18 SOC + 8.65 BD | 0.978 | 0.000 | 18 |
| | SICD = 31.209 − 0.464 SOC − 0.590 SWC | 0.907 | 0.000 | 18 |
| 10–20 | SIC = 17.61 − 0.35 SOC | 0.451 | 0.006 | 18 |
| | SICD = 21.555 − 4.742 RD | 0.754 | 0.000 | 15 |
| 20–30 | SIC = 21.12 − 0.40 SOC − 0.26 Sand | 0.706 | 0.001 | 18 |
| | SICD = 30.604 − 0.919 Sand | 0.816 | 0.000 | 18 |
| 30–50 | SIC = −3.11 + 0.23 Silt + 0.97 RD | 0.622 | 0.003 | 15 |
| 50–70 | SIC = 17.18 − 0.11 Sand | 0.604 | 0.001 | 18 |
| 70–100 | SIC = −18.13 + 4.27 pH | 0.402 | 0.011 | 18 |

† BD, bulk density; RD, root mass density; SOC, soil organic carbon; SWC, soil water content.

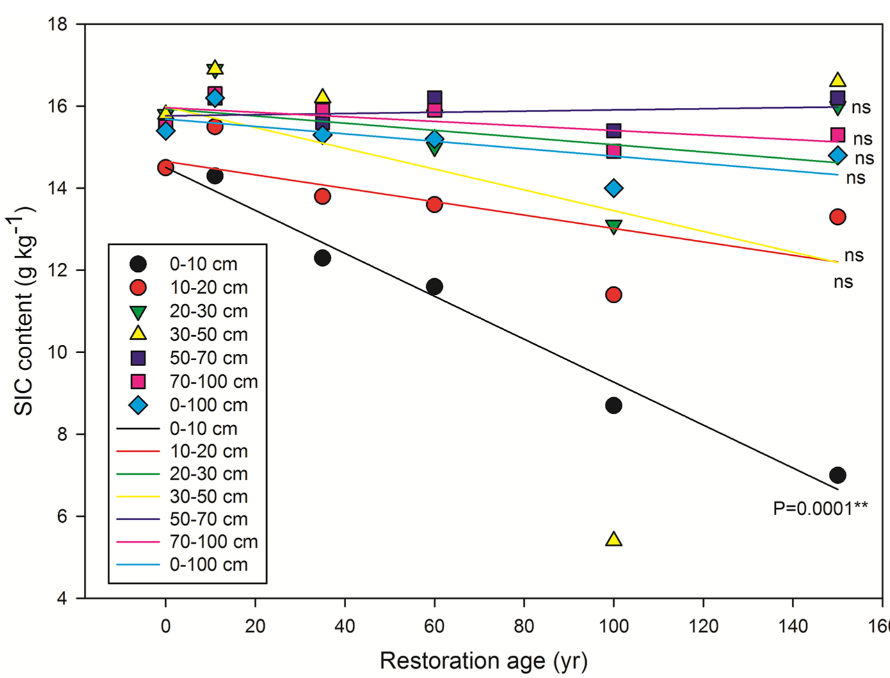


**Fig. 6. Linear regression analysis of the soil inorganic C content (SIC) in each layer across a 150-yr natural vegetation restoration chronosequence. ***P* < 0.01; ns, not significant (*P* > 0.05).**

tion of SIC to $CO_2$, which would ultimately be discharged to the atmosphere. In addition, the dissolved SIC would increase with vegetation age due to the increased moisture and decreased pH of the soil environment under restored vegetation. However, runoff would significantly decrease after vegetation restoration (Wang et al., 2011b). Thus, the decrease in surface runoff could reduce SIC loss across the vegetation restoration chronosequence. Therefore, the increasing $CO_2$ emissions induced by the decrease in soil $Ca^{2+}$ across the vegetation restoration chronosequence is likely to be the main reason for the decline in SICD in the 0- to 100-cm soil profile. This finding was supported by the results of Liu et al. (2014).

## Factors Influencing Changes in Soil Inorganic Carbon Content and Storage

According to the equilibrium reaction equation of carbonate dissolution–precipitation (Eq. [1] and [2]), soil moisture, $CO_2$ partial pressure, and $Ca^{2+}$ and $HCO_3^-$ concentrations are the direct factors that influence SIC. Thus, anything that has an effect on these factors would cause a change in SIC. The factors controlling SIC were different in different soil layers (Table 3), which reflected the complicated mechanisms and processes underpinning the formation and transformation of SIC. Soil organic C plays a very important role in determining the variation in SIC and had a significantly negative correlation with SIC in the top 50 cm of the profile (Table 3). The stepwise regression equations also confirmed the controlling function of SOC on SIC in the top 30 cm of the profile (Table 4). This was supported by the results of previous studies (Chang et al., 2012; Jelinski and Kucharik, 2009; Wang et al., 2013b). During vegetation restoration, not only did the aboveground litter and root biomass increase but the abundance and activity of soil microorganisms was also enhanced, which favored the accumulation of SOC (Deng et al., 2013; Wang et al., 2011a). The continuing increase in SOC would improve the soil structure, with a specific decrease in BD and pH and an increase in SWC (Table 2). Moreover, the greater SOC and biological activity under restored vegetation would lead to greater heterotrophic and autotrophic respiration, which results in higher $CO_2$ partial pressures (Chang et al., 2012). The accumulation of SOC under restored vegetation could induce an increase in carbonic and organic acid production, which reduces the availability of soil $Ca^{2+}$ through cation exchange in the soil (McLaughlin and Wimmer, 1999; Sartori et al., 2007). This would increase the dissolution and leaching of carbonate in the topsoil and decrease the SIC concentration, thereby reducing the SICD. However, the effects of SOC on the SIC decrease with soil depth and the correlation between SOC and SIC were insignificant below 50 cm.

For the SIC in the subsoil, the silt and sand content had a significant relationship with SIC in the 20–100 cm layers (Table 3). Stepwise regression also revealed the strong influence of the soil silt and sand content on SIC below 20 cm (Table 4). Studies have shown that soil texture plays an important role in the formation of PIC (Nordt et al., 2006; Rasmussen, 2006). The formation of PIC occurs more rapidly in coarse-textured soils than in fine-textured soils (Goddard et al., 2009). However, a significant positive correlation between SIC and the soil clay content was observed in desert biomes in Arizona, and the positive correlation between SIC and clay content indicated that these two properties co-vary as a function of soil development (Rasmussen, 2006). In our study, the soil clay content did not show any significant relationship with SIC in the profile due to its weak contribution to soil particle composition; however, the silt content

**Table 5. Correlations of soil organic C (SOC) and silt and sand contents with bulk density (BD), soil water content (SWC), pH, particle size distribution (sand, silt, and clay contents), and root mass density (RD) in different soil layers across a 150-yr vegetation restoration chronosequence.**

| Soil layer | Parameter | BD | SWC | pH | Clay | Silt | Sand | RD |
|---|---|---|---|---|---|---|---|---|
| cm | | | | | | | | |
| 0–10 | SOC | −0.872** | 0.500* | −0.334 | −0.233 | −0.405 | 0.439 | 0.640* |
| | silt | 0.284 | −0.051 | −0.030 | 0.137 | 1 | −0.962** | −0.626* |
| | sand | −0.363 | 0.173 | 0.094 | −0.403 | −0.962** | 1 | 0.673** |
| 10–20 | SOC | −0.532* | 0.167 | −0.518* | 0.272 | −0.282 | 0.163 | 0.569* |
| | silt | 0.447 | −0.114 | 0.338 | 0.097 | 1 | −0.940** | −0.819** |
| | sand | −0.342 | 0.178 | −0.379 | −0.429 | −0.940** | 1 | 0.842** |
| 20–30 | SOC | −0.164 | −0.445 | −0.446 | 0.231 | −0.392 | 0.328 | 0.548* |
| | silt | 0.605** | 0.653** | 0.516* | −0.220 | 1 | −0.952** | −0.639* |
| | sand | −0.658** | −0.682** | −0.491* | −0.090 | −0.952** | 1 | 0.626* |
| 30–50 | SOC | −0.031 | 0.100 | −0.600* | 0.096 | −0.369 | 0.355 | 0.165 |
| | silt | −0.262 | 0.289 | 0.695** | −0.233 | 1 | −0.970** | −0.768** |
| | sand | 0.202 | −0.370 | −0.775** | −0.010 | −0.970** | 1 | 0.816** |
| 50–70 | SOC | −0.040 | 0.148 | −0.044 | −0.574* | −0.017 | 0.192 | 0.082 |
| | silt | −0.043 | 0.208 | 0.810** | −0.182 | 1 | −0.954** | −0.595* |
| | sand | 0.031 | −0.257 | −0.780** | −0.121 | −0.954** | 1 | 0.534* |
| 70–100 | SOC | −0.050 | −0.130 | 0.328 | −0.446 | 0.168 | −0.008 | −0.234 |
| | silt | 0.115 | −0.600** | 0.830** | −0.386 | 1 | −0.932** | −0.256 |
| | sand | −0.062 | 0.560* | −0.739** | 0.027 | −0.932** | 1 | 0.182 |

* Correlation is significant at the 0.05 level.

** Correlation is significant at the 0.01 level.

had a significantly positive correlation with SIC, while the sand content had the inverse correlation (Table 2). Coarse-textured soils have a high porosity and permeability, which results in greater hydraulic conductivity than fine-textured soils (Woods and Balfour, 2010). The higher hydraulic conductivity in coarse-textured soils favors the dissolution and leaching of carbonate. In contrast, fine-textured soils, with a lower hydraulic conductivity, are prone to retain carbonate in the soil profile.

In addition, BD, RD, SWC, and pH also had different degrees of correlation with SIC throughout the soil profile (Table 3). However, most of these properties co-varied with SOC and the silt and sand contents in the corresponding soil layers (Table 5). Thus, SOC and the silt and sand contents could be treated as comprehensive indexes to characterize the status of SIC during vegetation restoration. In summary, vegetation restoration had a strong effect on SIC in the topsoil, with a significant negative correlation between SOC and SIC, while the genetic soil features determined the SIC in the subsoil, with a significant positive correlation with silt and SIC and a significant negative correlation between the sand content and SIC.

### Effect of Soil Inorganic Carbon Storage on Soil Total Carbon Storage

Many previous researchers have reported that vegetation restoration had a significant positive effect on SOCD (Guo and Gifford, 2002; Hernandez-Ramirez et al., 2011; Deng et al., 2013). This result was supported by our present study. However, our study also found a decreasing trend of SICD in the 0- to 10- and 0- to 100-cm soil layers, which offset the effect of the SOCD accumulation and led to an insignificant change in the STCD across the vegetation restoration chronosequence (Fig. 4). Our observations are in agreement with previous studies that also reported no significant difference in STCD among farmland and vegetation restoration sites (Jelinski and Kucharik, 2009; Chang et al., 2012; Liu et al., 2014). There could be two reasons to explain the insignificant change in the STCD across the chronosequence. First, the significant negative relationship between SOC and SIC above the 50-cm soil depth indicated that the increases in SOC would be balanced by the decreases in SIC across the vegetation restoration chronosequence (Table 3). Second, because SIC was the dominant form below 50 cm and remained unchanged across the restoration chronosequence, there was little effect on the total soil C pool (Fig. 2). Therefore, no difference was detected in the STCD across the chronosequence; there was just a redistribution of the total soil C pool from SIC- to SOC-dominated forms. All in all, many previous studies have looked only at SOC or SIC but an evaluation of both is needed for calcareous soils to determine the total impact of restoration on the soil C dynamics.

## CONCLUSIONS

Vegetation restoration had a strong effect on both the distribution and amount of SICD. Compared with farmland, a significant decline was observed in the amount of SICD in the 0- to 10-cm soil depth along a 150-yr natural vegetation restoration chronosequence. However, the amount of SICD remained basically unchanged in the subsoils (10–100 cm). The contribution of SICD to the soil C pool clearly increased with soil depth; however, it decreased from 65.2 to 43.3% during the 150-yr restoration chronosequence. The total soil C storage did not change significantly with vegetation restoration due to the offset effect of an SOC increase and an SIC decrease and also because of the buffering action of a large and stable SIC pool in the subsoil. The controlling factors on SIC were different in the top- and subsoils along the chronosequence. In the top 30-cm soil layers, SOC showed a significant correlation with SIC; however, in soils below 30 cm, the sand and silt content and pH were better predictors of SIC. Therefore, we can conclude that the variations in SOC induced by vegetation restoration are the main driving force of SIC changes in the topsoil, while the genetic soil features (i.e., sand and silt content) were the controlling factors that determined the amount of SIC in the subsoil.

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