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ORIGINAL PAPER

High organic carbon stock in a karstic soil of the Middle-European Forest Province persists after centuries-long agroforestry management

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Abstract Little is known on soil organic carbon (SOC) stocks in karst areas worldwide, although many of them have seen long-term application of agroforestry systems with a potential for carbon sequestration. Therefore, our study aimed to assess landscape-level SOC concentration and stock in the Silica Plateau, a part of the Slovak Karst Biosphere Reserve located in the Western Carpathians (Slovakia) with a centuries-long agroforestry record. The most represented local soil units are Chromi-Rendzic Leptosols and Chromic Cambisols with clayey loam texture, C/N ratio 9–12, and pH_{H_2O} 6.6–6.2 in their organo-mineral surface horizons. Mull surface humus form prevails under mixed forest stands dominated by hornbeam (*Carpinus betulus* L.), oak (*Quercus petraea* L.), and beech (*Fagus sylvatica* L.). A total of 2,700 soil samples were collected from 150 soil pits. Both SOC concentrations and stocks were determined for the 0–60 cm mineral soil layer. Soil stoniness was accounted for by means of electrical resistivity tomography. According to the analysis of covariance, cropland SOC concentration (0.026 g g^{-1}) is significantly lower compared to forestland (0.040 g g^{-1}) and pastureland (0.041 g g^{-1}) ($P < 0.01$). During the period of 130 years after forest clearing, cropland SOC stock has been reduced at an exponential decay rate of ca 0.002 $year^{-1}$, while the SOC stock in pastureland has increased following land use change from cropland by approximately 30% during the same period of time. Irrespective of land use history, overall SOC stock is high reaching on average 207.4 Mg ha^{-1}, out of which 66% are stored within 0–30 cm and 34% within 30–60 cm soil layers.

Keywords Calcareous soils · Soil organic carbon · Soil stoniness · Land use · Electrical resistivity tomography

Introduction

Soil organic carbon (SOC) is essential in determining the physical and chemical properties of soils, as well as in the sustenance of primary production in terrestrial ecosystems and the process of trapping atmospheric CO_2 (Santore et al. 1995; Post and Kwon 2000). Global soil organic matter exceeds the sum of the atmospheric and biotic pools (Bouwman and Leemans 1995; Schlesinger 1997; Lal 2004). Carbon stock in soils depends on geological substrate, precipitation and temperature (Turrión et al. 2009; Luo and Zhou 2006; Davidson and Janssens 2006), terrain topography and microtopography (Bergstrom et al. 2001a; Stoeckel and Miller-Goodman 2001; Eglin et al. 2008),

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land use and soil tillage (Baker et al. 2007), and other factors, such as soil depth under consideration (Jobbágy and Jackson 2000). The relative importance of these factors is scale dependent (e.g. Bergstrom et al. 2001b; Corstanje et al. 2007). Although carbon bound in the biomass may exceed the amount of SOC stored within surface humus and topsoil (0–50 cm) of European forest, the overall SOC content is still very significant, ranging from ca 60 to 230 Mg ha^{-1} (Bauer et al. 2000).

The carbon balance results from C fluxes driven by photosynthesis and respiration, the latter process being dominated by root and microbial soil respiration (Valentini 2003). Generally, soil organic matter (SOM) mean residence time depends on SOM quality and the level of its physical and chemical protection (e.g. Jenkinson and Rayner 1977; Stevenson 1994; Sollins et al. 1996; Zaccone et al. 2011). Consequently, ca 1,500 Pg of organic carbon in the upper meter of mineral soils (Jobbágy and Jackson 2000) consists of several pools with characteristic turnover times (Parton et al. 1987). Thus, approx. 200–300 Pg C in SOM exist in forms that recycle during a century or less (Schimel 1995; Potter and Klooster 1997), while the remainder is stable on time scales of centuries to millennia (Trumbore 2000). Harrison et al. (2000) reported SOM mean resident times ranging from decades to several 100 years in European beech and spruce forests. Importantly, significant proportion of SOC found bellow 10 cm was locked up in stable forms, illustrating the importance of forest soils for the terrestrial carbon cycling in Europe in comparison with the more vulnerable organic carbon stored in agricultural soils and drained peats. The annual SOC loss caused by changes in land use can be considerable, although large variation exists in different parts of the world (Bouwman and Leemans 1995). For example, while forest management has little effect on SOC (Johnson and Curtis 2001), soils of the world's agroecosystems are often depleted of their SOC pool by 25–75% depending on climate, soil type, and historic management (Lal 2011). On the other hand, Montagnini and Nair (2004) estimated average carbon storage by agroforestry practices at 63 Mg ha^{-1} in temperate regions, suggesting that agroforestry systems with perennial crops may be important carbon sinks. Also, agroforestry systems involve trees or shrubs, agricultural crop, and possibly pasture as their main components (Mosquera-Losada et al. 2009), thus preserving their attractiveness from ecological and socio-economic points of view (Franco et al. 2001).

However, few landscape-level SOC stock data sets are available for agroforestry systems in karst areas, despite their high representation in Europe and worldwide (e.g. Rivera et al. 2000; Wang et al. 2004; Vidrih et al. 2009). In karst areas, SOC stock calculations heavily depend on visual stoniness assessment in the field, which may be uncertain. For example, Wirth et al. (2004) established that soil stoniness was overestimated by 40%, which would imply an underestimation of SOC stock. Therefore, our study aimed to quantitatively assess SOC stock and variability within a temperate zone agroforestry system, as well as to estimate the impact of past and present land use on SOC concentration in a karst landscape, using electrical resistivity tomography as a non-destructive auxiliary approach.

Materials and methods

Description of study area

Silica Plateau is located within the Slovak Karst Biosphere Reserve in the Western Carpathians, Slovakia. Our experiment was conducted within the area extending between 48°33′51.22″ and 48°36′26.63″N and 20°29′33.81″ and 20°33′03.30″E. The plateau is composed of impermeable Lower Triassic sediments overlain by the Middle Triassic limestone-dolomite complex that contain ca 2% of insoluble remains, mainly illite, montmorillonite, feldspars, quartz, and mica (Šály 1978; Mello et al. 1996). Area's mean altitude is approximately 600 m a. s. l. Silica Plateau falls to the warm, moderately humid region with cold winter (Lapin et al. 2002) with the mean annual air temperature 5.7–8.5°C and annual average precipitation 630–990 mm (400–595 mm during the vegetation period). Local soils are mostly represented by Rendzic Leptosols, Chromi-Rendzic Leptosols, and Chromic Cambisols (acc. to WRB; FAO 2006), the latter being bound to accumulation positions. Both Rendzic and Chromi-Rendzic Leptosols have $pH_{H_2O} \cong 6.6$ within 0–40 cm. In Chromic Cambisols, $pH_{H_2O} \cong 6.2$. The C/N ratio in soils ranges 9–12. Average cation exchange capacity in the humus-rich topsoil reaches ca 450–500 mval kg^{-1} and it decreases with depth. Concerned soil units fall into the clayey loam textural class. The dominant clay mineral phase in the clay fraction of the Silica soil is illitic material. Soils depth ranges from 20 to 40 cm on ridges to more than 1 m in accumulation positions (Kobza 1994; Šály 1994; Miko et al. 2003).

According to field survey, current forestland is covered by mixed stands of common hornbeam (*Carpinus betulus* L., 50%), sessile oak (*Quercus petraea* L., 30%), and European beech (*Fagus sylvatica* L., 20%) that fall into several alliances and sub-alliances, mainly *Carici pilosae-Carpinenion betuli*, *Quercion pubescentis-petraea*, and *Cephalantero-Fagenion*. Tree litter of the respective broadleaved mixture has a comparatively low C/N ratio of ca 44 due to the dominance of hornbeam tree litter (C/N ratio ≅ 34) (Bublinec 1994; Hättenschwiler and Gasser 2005). Occurrence of conifers (*Picea abies* K.; tree litter

C:/N ratio $\cong$ 50) is very limited (1–2%) and confined to sink holes in the northern part of the plateau. The shrubland vegetation belongs to several alliances, mainly the xerophytic *Prunion fruticosae*, represented by *Prunus* spp., mainly *Prunus fruticosa* with tree litter C/N ratio $\cong$ 78 (Lorenz et al. 2004). *Prunus* spp. are accompanied by other shrubs, notably *Rosa galica* L., *Crataegus monogyna* L., and *Berberis vulgaris* L. Pastureland is covered by perennial grasses from the alliance *Cynosurion cristati* Tüxen 1947, for example, *Festuca rupicola* Heuff., *Briza media* L., *Lolium perenne* L., and a smaller portion of legumes (*Trifolium repens* L.). Avg. C/N ratio of the grassland vegetation litter (approx. 45) is an estimate adopted from Hegyhátsál (Hungary), botanically related and geographically close locality (Tomelleri 2007; Janišová et al. 2007). The pastureland provides sustenance for roaming herds of cattle and sheep. Their fresh feces and urine have C/N ratio about 23 and 1, respectively (Kirchmann 1985). Cropland has been used to produce cereal grains, mainly rye (*Secale cereale* L.) and common oat (*Avena sativa* L.), whose C/N ratio in the crop residues ranges 75–100 (Axmann et al. 1990), as well as alfalfa (*Medicago sativa* L.) with a much lower crop residues C/N ratio $\cong$ 13 (Troeh and Thompson 1993). Average annual addition of mineral N through fertilization has been ca 20–100 kg ha^{-1} year^{-1} since 1967 (Rozložník and Karasová 1994).

Land use history

The land cover mosaic has been changing during past centuries, as documented by map surveys carried out in 1780, 1850, and 1880, as well as modern-time aerial photography data (1950) and current CORINE Land Cover data (2010). The spatio-temporal changes reflect several turning points, starting with the Great Turkish War that played out mainly during the seventeenth century's 2nd half. The war caused significant decline in the population density and land abandonment on the concerned territory (Wessely 1973). As a result, large tracts of land survived as forests or they were re-claimed by forests (Fig. 1a). By the eighteenth century's 2nd half, new clearings appeared amidst continuous forests covering Silica Plateau owing to population recovery and the return of refugees to the former war zone. Larger clearings were turned into cropland and pastureland. During the 1st half of the nineteenth century, the woodland cover was further reduced and

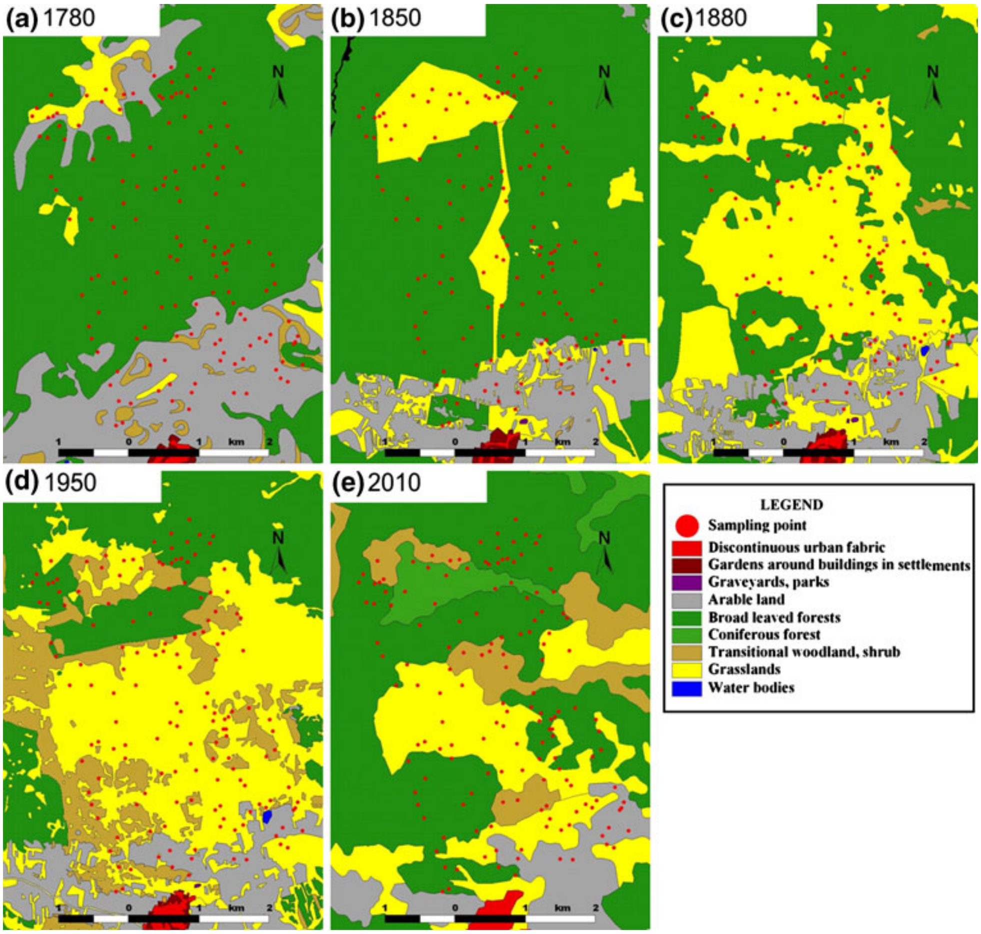

Fig. 1 Maps reproducing land cover changes on the Silica Plateau between 1780 and 2010

converted into pastureland. By and throughout the nineteenth century's 2nd half, less profitable soil tillage in the more remote parts of Silica plateau gave way to the process of secondary succession and forest restoration, and large portion of cropland and forestland were converted to pastureland. As a result, the area of pastureland and shrubland had increased by 1950. Area-wise land cover developments are captured by Table 1. For the purpose of further analysis, we included strongly fragmented shrubland into forestland because of uncertainties inherent to historical map surveys and the tree species demography across the invasion front. This front can extend ca 200 m from forest boundary toward pastureland (Dovčiak et al. 2008).

Soil survey and sampling scheme

One hundred and fifty sampling points were obtained using random numbers generator (Mathematica 8.1; Wolfram Research, Inc., Champaign, IL) and projected on a rectangle area 3 km wide and 5 km long, designated within the Silica Plateau. At each point, soil depth and stoniness were surveyed by electrical resistivity (ER) tomography (ARES, GF system, Brno, Czech Republic) and subsequent imaging of soil/bedrock interface. ER (Ω-m) was acquired by an arrangement of 32 electrodes with a spacing of 0.8 m according to Rey et al. (2006). The processing and reconstruction of 2-D resitivity profiles was implemented in RES2DINV program (Geotomo Software, Gelugor, Malaysia). Coarse fragments content (hereinafter also referred to as stoniness), needed for the establishment of the corresponding calibration relationship with ER, was determined on six soil profiles, which were prepared by a specially designed sledge-hammer-driven spade. The profiles were photographed and the photographs were corrected for geometric distortion. Area occupied by rocks was determined by image analysis, that is, color-based segmentation using the minimum average pixel classification error (Kittler and Illingworth 1986) according to the routine in Mathematica Digital Image Processing 1.1 (Wolfram Research, Inc., Champaign, IL). Stoniness within concerned soil depth (0–60 cm) was determined as the sum of: (1) the relative volume of stones (20–200 mm) and boulders (>200 mm), which is proportional to their relative area on the profile wall in soil pits (Folk 1951; Alexander 1982), previously established by the image analysis; (2) the volume of gravel (2–20 mm). It was measured in undisturbed soil samples (200 cm^3), also collected for the determination of fine earth bulk density. They were taken from fifteen profiles representing all land covers (cropland, pastureland, and forestland) at 6 depths (10, 20, …, 60 cm). Mineral soil samples for SOC determination were taken from profiles exposed by pick, spade and shovel on all 150 points. Samples were carved out from six layers (0–10, 10–20, …, 50–60 cm) by means of knife with triple replication along vertical lines, 20 cm apart. Samples weighing ca 500 g were collected in plastic bags. A total of 2,700 samples were collected.

Table 1 Land use changes in the Silica Plateau according to historical, recent, and current surveys

Land cover	Area according to land use (%)				
	1780	1850	1880	1950	2010
Forestland	64.0	68.0	30.0	19.3	54.0
Shrubland	3.3	0.7	0.0	16.7	12.0
Pastureland	6.0	20.7	59.3	56.7	26.0
Cropland	26.7	10.7	10.7	7.3	8.0

Soil analyses

Soil samples were air-dried, ground, and passed through a 2-mm mesh sieve. Visible plant residues, gravel, and stones were removed manually with a pair of tweezers, while the small root fragments were removed by electrostatically charged stick (Kuzyakov et al. 2001). The C and N contents in the fine earth (<2 mm) were determined by Vario MACRO Elemental Analyzer (CNS version, Elementar GmbH, Hanau, Germany), which employs the dry combustion (DC) method. Because elemental analyzer provides total carbon contents, inorganic C content was measured separately for each sample by a volumetric device (Fiala et al. 1999) and subtracted from the total carbon in order to obtain SOC mass concentration [$\mathrm{M\ M^{-1}}$]. The commensurability of our SOC concentration analysis with a different method under site-specific conditions was assessed by comparing our preliminary forestland SOC concentration values, determined by DC method ($\mathrm{SOCC_{DC}}$, $n_1 = 12$), with the full set ($n_2 = 12$) of forestland SOC concentrations ($\mathrm{SOCC_{WO}}$), previously obtained in the area of interest and available in Šály (1994). $\mathrm{SOCC_{WO}}$ were determined by a wet oxidation (WO) titrimetrical method according to Tyurin (1931; also see Kononova 1966). To assess the difference between the two sample means (avg. $\mathrm{SOCC_{DC}} = 0.40$, SD: 0.02; avg. $\mathrm{SOCC_{WO}} = 0.36$, SD: 0.02), the t test (Sokal and Rohlf 1995) was applied to both data sets with and without the application of general conversion coefficient between DC and WO methods compiled by Jankauskas et al. (2006). Based on the test results, alternative hypothesis about significant difference between sample means was rejected ($P = 0.83$–0.57).

Because we knew bulk density and volume of the soil (i.e., fine earth plus soil pores, without coarse fragments with diameter > 2 mm), SOC stock was computed by summing up the C content in all six 10 cm layers at each sampling point according to

$$\mathrm{SOCS} = \sum_{i=1}^{6} \mathrm{BD}_i \times \mathrm{SOCC}_i \times d_i \times (1 - \mathrm{cf}_i), \qquad (1)$$

where SOCS, BD_i, SOCC_i, d_i, and cf_i, are, respectively, SOC stock [M L^{-2}], soil bulk density [M L^{-3}], SOC concentration [M M^{-1}] in layer i, ith layer thickness [L], volumetric fraction of coarse fragments in layer i [L^3 L^{-3}]. Similar relationships were used by Rodríguez-Murillo (2001), Stevens and van Wesemael (2008) and others.

Data evaluation

Differences among average soil bulk densities according to land cover were assessed by ANOVA, including Tukey-HSD post hoc comparison test. The analysis of covariance (ANCOVA) (Sokal and Rohlf 1995) was used to evaluate the effect of historical and current land covers on SOC concentration, in that land cover and stoniness entered the analysis as categorical (predictor) and continuous (confounding) variables, respectively. Average SOC stock (Mg ha^{-1}) was calculated for the whole area of interest based on all sampling points and whole soil profiles (0–60 cm). All analyses were performed within Statistica 9 (StaSoft, Inc., Tulsa, USA). Furthermore, we used the method of chronosequences by calculating the rates of SOC stock depletion or recovery following two land use changes (LUC) (1) from forestland to cropland and (2) from cropland to pastureland for points featuring similar soil depths and stoniness. Because two military surveys (1850 and 1880) were separated by just 30 years, we omitted the 1850 map and data, as capturing transition rather than at least temporary steady state in further analyses.

Results and discussion

Soil depth, stoniness, and bulk density

Soils depth and stoniness were instrumental to our study. Both variables were surveyed using the relationship between ER and stoniness. The relationship was fitted by a decadic logarithm function (Fig. 2) with corresponding root-mean-square error. Our calibration corresponds well to data provided by Marescot (2006), who gave ER about 1,000 Ω-m and above for limestone materials. Examples of soil profiles featuring near-zero and high skeleton contents are given in Fig. 3.

Leaning on the calibration, our ER tomography survey revealed coarse fragments content across the entire area (Fig. 4). Such an irregular distribution is typical of karstic surfaces that include both erosive and accumulation positions such as rocky slopes and ridges, grike fields, dolines, sinkholes, and other characteristic features. The comparison between Figs. 1a–e and 4 illustrates a correspondence between extensively (forestland and pastureland) or intensively (cropland) managed land covers on one hand, and stony or finely grained soil patches on the other hand. Such overlap patterns are common to land use and forest cover histories in Europe (Pichler et al. 2011). Obviously, the non-destructive survey provided a sound and less labor and time-consuming alternative to quantitative soil pits method, which could not be applied due to the large volume of stones and boulders.

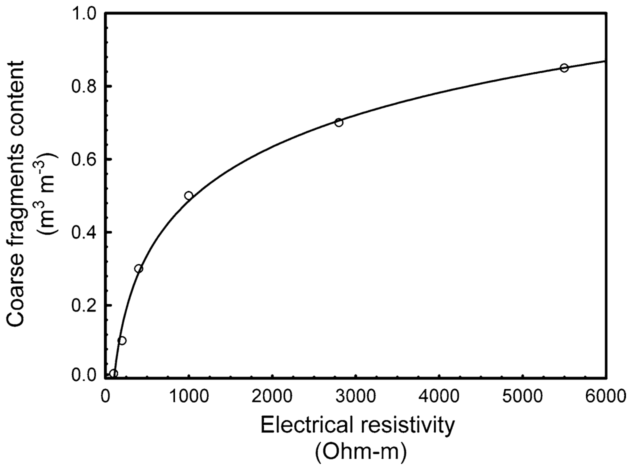

Fig. 2 Coarse fragments content (CFC) versus electrical resistivity (ER) fitted by a decadic logarithm function (CFC = $-0.99 + 0.49 \log_{10}$ ER; root mean square error: 1.9×10^{-2})

Soil bulk density for distinct land covers is given in Table 2. The bulk density increases with soil depth and from forestland to cropland due to higher SOC concentration in the topsoil and in forest soils. Also, coarse fragments below the soil surface, abundantly present in forestland and, to a lesser degree, in pastureland, support the existing soil structure and reduce soil compactibility (Poesen and Lavee 1994).

SOC stock and variability

Our SOC stock calculation was based on a calibrated stoniness–ER relationship. The stock within the 15 km^2 area of interest averaged 207.4, 136.8, and 70.7 Mg ha^{-1} in 0–60, 0–30, and 30–60 cm soil layers, respectively. This high amount of organic carbon (e.g., 103.3 Mg ha^{-1} within 0–20 cm layer) fell slightly above the 75% CI for the mean SOC stock calculated for Rendzic Leptosols (also within 0–20 cm), but well below the maximum value 126.3 Mg ha^{-1} (Baritz et al. 2010). This is probably due to the role of Ca^{2+} in the cation bridging of organic colloids, condensation and stabilization of organic matter (Oades

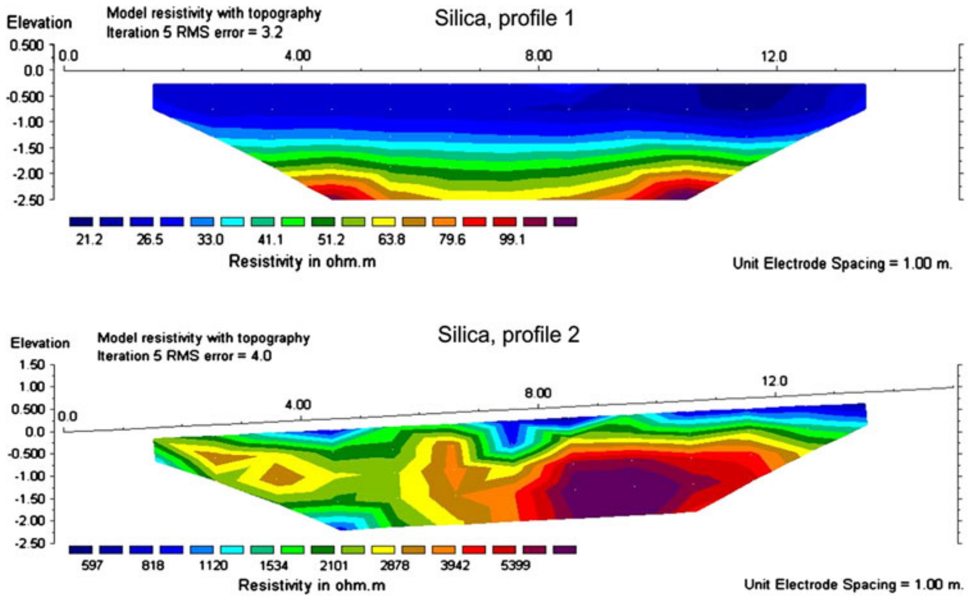

Fig. 3 Laterally homogeneous, low-electrical resistivity (ER) profile taken in a karst "polje" position (profile 1) and heterogeneous, high-ER profile taken in ridge position (profile 2)

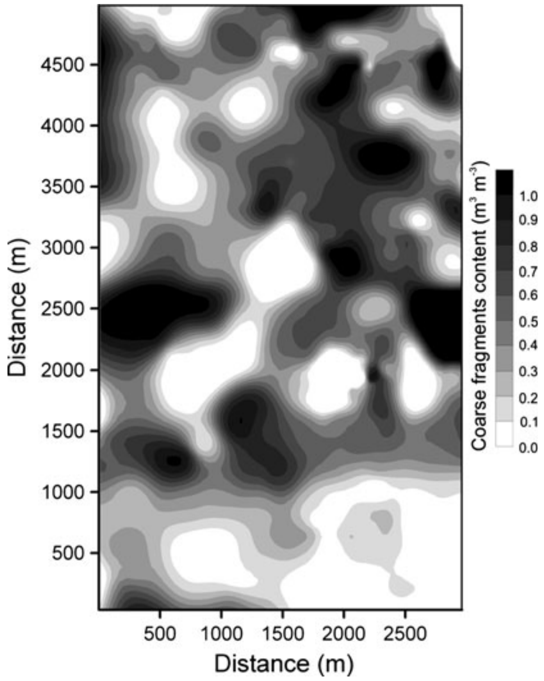

Fig. 4 Soil stoniness distribution on the Silica Plateau produced by the modified Shepard's method (Franke and Nielson 1980) from all 150 points

Table 2 Average soil bulk densities (BD) and corresponding standard deviations calculated from 5 undisturbed samples (200 cm^3) for each land cover

Soil depth (cm)	Forestland		Pastureland		Cropland	
	BD (Mg m^{-3})	SD	BD (Mg m^{-3})	SD	BD (Mg m^{-3})	SD
0–10	1.21^{a}	0.03	$1.38^{a,b}$	0.05	1.58^{b}	0.14
10–20	1.27^{a}	0.03	1.39^{b}	0.03	1.60^{c}	0.02
20–30	1.33^{a}	0.05	1.41^{a}	0.02	1.62^{b}	0.09
30–40	1.39^{a}	0.04	1.42^{a}	0.02	1.64^{b}	0.09
40–50	1.44^{a}	0.03	1.43^{a}	0.04	1.66^{b}	0.10
50–60	1.51^{a}	0.07	1.44^{a}	0.11	1.69^{b}	0.10

Distinct superscript letters associated with BDs for the same depth indicate significant differences ($P < 0.05$) among BD means, as established by ANOVA and Tukey-HSD post hoc comparison test

1988), formation of thin carbonate coatings on particulate organic matter (Duchaufour 1976), and SOM physical protection from decomposition through cementation of soil aggregates by $CaCO_3$ (Oyonarte et al. 1994). Average SOC concentration in Silica soils was 0.039 g g^{-1} (range: 0.015–0.077 g g^{-1}), 0.049 g g^{-1} (range: 0.019–0.105 g g^{-1}), and 0.027 g g^{-1} (range: 0.006–0.070 g g^{-1}) for 0–60, 0–30, and 30–60 cm, respectively. These are comparatively high values, but lower than 0.050–0.150 g g^{-1} as found in tropical semiarid calcareous soils of northwestern Yucatán, Mexico (Shang and Tiessen 2003). However, the latter work provides certain analogy to our results, in that higher SOC concentrations were found in shallow black lithosols surrounding rock outcrops, while lower SOC concentration was usually established in deeper red rendzinas at slightly lower relief. Also, despite increasing soil respiration rates, greater biological activity in limed soils leads to plant C inputs being processed and incorporated into resistant soil organo-mineral pools (Fornara et al. 2011). However, the variation of SOC stock within the relatively small study area was considerable with a SD of 85.7 Mg ha^{-1}. This formidable variability, resulting from highly variable SOC concentration, rock fragments content, bulk density, and land cover could explain the rareness of SOC stock estimates from karstic soils.

SOC concentration in the fine earth according to land cover

To assess SOC dependence on land cover, we preferred to analyze SOC concentration instead of SOC stock, because calculation of the latter variable involves strongly fluctuating stoniness and, to a lesser degree, soil bulk density. Intrinsic connection between SOC concentration and stoniness is implicitly assumed by various authors, but it was directly suggested only by few, for example, Schaetzl (1991) and Tate et al. (2005). In addition, stoniness and land use were linked through the process of land allocation for specific uses. In fact, dispersion of SOC concentration in cropland was limited because soil tillage was confined to the least stony soils. This is obvious from Fig. 5, which reproduces the location and distribution of SOC concentration according to land cover. Existing collinearity among SOC concentration, land cover, and stoniness (Figs. 5, 6) was controlled for and handled by ANCOVA, in which land cover, stoniness, and SOC concentration represented categorical variable (fixed factor), continuous (confounding) variable, and dependent variable, respectively. The results of ANCOVA, as performed for land cover mosaics representative of four different periods of time, are given in Table 3.

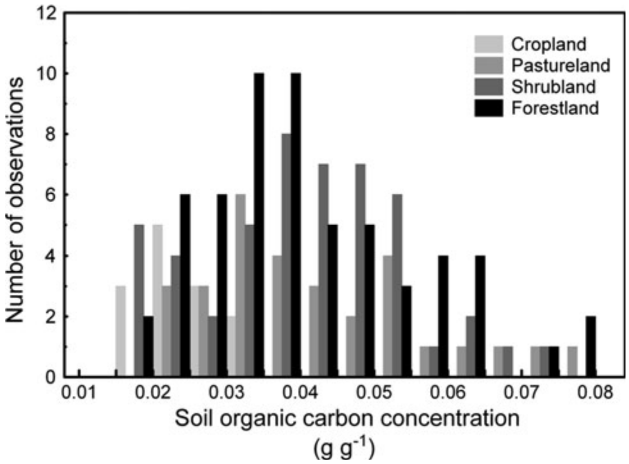

Fig. 5 Histogram of depth-averaged soil organic carbon concentrations in the fine earth according to land cover

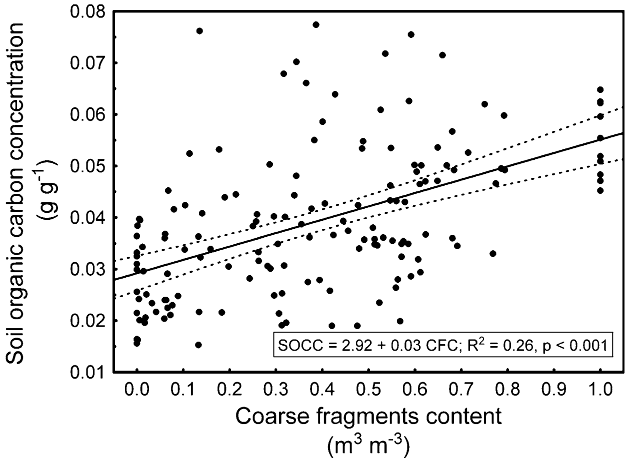

Fig. 6 Linear relationship (*solid line*) between soil organic carbon concentration in the fine earth (SOCC) and soil coarse fragments content (CFC); dashed lines show 95% CI

The analysis of covariance suggests that SOC concentration imprint of current (2010) land cover is significant in that SOC concentration in cropland (0.026 g g^{-1}) is lower compared to forestland (0.040 g g^{-1}) and pastureland (0.041 g g^{-1}) ($P < 0.01$ according to ANCOVA, including the Tukey-HSD post hoc test). Šamonil (2007) reported similar SOC concentrations (ca 0.04 g g^{-1}) in Rendzic Leptosols under forest fragments covering parts of the Bohemian Karst. In our study, the effect of the most recent land cover was also detected in the subsoil (30–60 cm) ($P < 0.05$). Although similar effect of the previously recorded land cover (1950) probably still persists ($P < 0.13$), the probability fell sharply for older periods of time ($P < 0.24$, 0.47) owing mainly to more frequent land use swaps between cropland on the one hand and forestland or pastureland on the other hand. Thus, expectedly, the supposed effects of land use on SOC concentrations, for example, due to soil tillage in the NW part of the area of interest back in 1780 (Fig. 1a), have become intractable due to SOC recovery following secondary forest succession or grazing. Because SOC concentrations under forestland and pastureland have been almost identical in the Silica soil, our investigation confirms that the general assumption about the capacity of both forestland and pastureland to retain more SOC compared with cropped land use (e.g. Martens et al. 2003) is also valid for karstic soils. Stark contrast between cropland and the two remaining land covers also shows in the vertical distribution of SOC concentration (Fig. 7), which was satisfactorily fitted by an

Table 3 Analysis of covariance of the soil organic carbon concentration in the fine earth according to land cover (fixed factor) and stoniness (confounding variable); *df*—degrees of freedom, η^2—partial eta-squared

Period of time	Factor	*df*	*F*-ratio	*P*	η^2
1780	Land cover	2	0.76	0.47	0.01
	Stoniness	1	36.99	0.00	0.20
1870	Land cover	2	1.45	0.24	0.02
	Stoniness	1	37.04	0.00	0.20
1950	Land cover	2	2.11	0.13	0.03
	Stoniness	1	34.10	0.00	0.19
2010	Land cover	2	5.12	0.01	0.07
	Stoniness	1	33.14	0.00	0.18

$\eta^2 = SS_{treatment}/(SS_{treatment} + SS_{error})$; SS—sum of squares exponential function, usually applied in similar studies (e.g. Bernoux et al. 1998).

Soil organic carbon stock depletion, recovery, and reallocation

To achieve a more reliable and conservative comparison between SOC stocks under certain land cover histories, we only considered SOC stocks within 0–60 cm for points with stoniness → 0. Respective SOC stocks under these constraints are given in Table 4.

The data provide an opportunity to assess the rate of SOC stock depletion or recovery due to forestland → cropland and cropland → pastureland LUCs, respectively, during the last 130 years. Since the LUC (1880), the respective Silica cropland soil has lost ca 25% of its initial SOC stock, which is less than an average of 32% after only 23 years, based on several studies evaluated by Poeplau et al. (2011). This was probably owing to the high initial SOC stock, and its effective clay and silt protection in the Silica soil, but also because available studies were mostly concerned only with to the topsoil (approx. 0–30 cm), where the relative loss was probably more pronounced and faster. Although statistical significance could not be established due to small number of observations, we can use exponential decay model (Olson 1963) to assess the rate of SOC stock depletion between land use histories 1 and 2 (Table 4):

$$\frac{\mathrm{dSOCS}_t}{\mathrm{d}t} = \lambda \mathrm{SOCS}_t \tag{2}$$

$$\mathrm{SOCS}_t = \mathrm{SOCS}_0 \mathrm{e}^{\lambda t} \tag{3}$$

$$t\lambda = \ln\left(\frac{\mathrm{SOCS}_t}{\mathrm{SOCS}_0}\right) \tag{4}$$

In (2–4), SOCS_t represents SOC stock at a time t counted from forestland → cropland LUC in 1880, λ is the exponential rate of decay, and SOCS_0 is the initial SOC stock (land use history No. 1 acc. to Table 4). From Table 4 and (4), we obtain $\lambda = 0.002\ \mathrm{year}^{-1}$, which is at the bottom of the range compiled by Tiessen et al. (1982). The reasons for tillage-induced SOC depletion include reduced litter inputs, increased soil aeration, and the loss of silt and clay protected C (Six et al. 2002). Also, the input of nutrients from crop residues and fertilization affected the soil biogeochemical cycle. Average C/N ratio in cropland soil (10) was significantly lower ($P < 0.05$) than in forestland and pastureland soils (12). Although the inputs of the organic C and N strongly contributed to this difference, the particular role of N in SOC dynamics depends on N input rate, initial SOC content, and lignin input to soils. Under low to moderate N input, for example, 40–100 kg N ha^{-1} year^{-1}, SOC concentration slightly increased or remained constant during several years of fertilization (e.g. Nyborg et al. 1995; Dijkstra et al. 2004; Šimon 2008). As opposed to that, fertilization rates beyond crop N requirements promotes SOC decline (Khan et al. 2007). Given relatively low to moderate amounts of mineral N added to the Silica Plateau cropland soils (20–100 kg ha^{-1} year^{-1}), we speculate that fertilization did not play dominant role in their SOC dynamics.

Table 4 also demonstrates SOC stock recovery after LUC from cropland to pastureland. The observed, moderate rise in SOC stock was probably sustained and affected by multiple factors. First, above- and below-ground litter, as well livestock feces may provide sufficient amount of C to maintain SOC stock in pastureland on levels similar to forestland (Takahashi et al. 2007). Besides, grazing normally leads to greater root allocation of C in the perennial plants (e.g. Briske et al. 1996; Stewart and Metherell 1999). Finally, SOC accumulation in pastureland may be supported by comparatively high resistance of humic acids from permanent meadows soils to microbial degradation reported by Filip and Tesařová (2005). On the other hand, SOC stock rise in pastureland was slow: it increased by only ca 30% during 130 years. It means that the corresponding rate of recovery was only about one-fourth of the average rate reported by Poeplau et al. (2011). Although SOC stock in pastureland has eventually reached forestland levels, it is

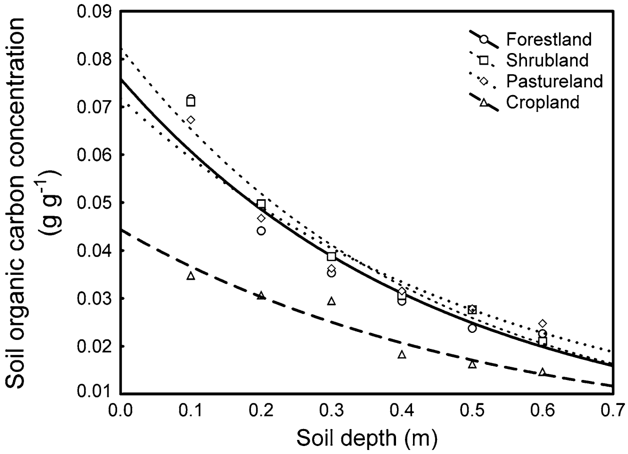

Fig. 7 Exponential function fitted on soil organic concentration in the fine earth according to land cover and soil depth

Table 4 Soil organic carbon stock in the fine earth within 0–60 cm at points with similar soil depth (±60 cm) and stoniness (0.0–0.1 $\mathrm{m}^3\ \mathrm{m}^{-3}$)

Land use history no.	Chronosequence 1780, 1880, 1950, 2010 (F-forestland, C-cropland, P-pastureland)	Number of points	SOC stock ($\mathrm{Mg\ ha}^{-1}$)
1	F_{1780}, F_{1880}, F_{1950}, F_{2010}	5	307.4
2	F_{1780}, C_{1880}, C_{1950}, C_{2010}	3	230.7
3	C_{1780}, C_{1880}, C_{1950}, C_{2010}	6	239.7
4	C_{1780}, P_{1880}, P_{1950}, P_{2010}	6	313.5

also possible that its slower-than-expected rise and the low rate of SOC depletion in cropland have been mutually related through erosive losses and subsequent accumulation. Erosion was reported to play important role in local SOC depletion (e.g.Voroney et al. 1981; Kimble et al. 2001; Schwanghart and Jarmer 2011). In our case, a portion of the C-rich topsoil material from grazed patches in the ridge and slope positions was likely transported to terrain depressions, mainly sinkholes and dolines. Thus, erosive losses and accumulation within the Silica Plateau affect not only soil depth (Miko et al. 2003), but most likely also contribute to the observed spatio-temporal variability of SOC stock.

Conclusions

The application of electrical resistivity tomography, leaning on the electrical resistivity—stoniness calibration, allowed us to acquire a detailed overview of the coarse fragments distribution over a 15 km^2 area located on the top of the Silica Plateau in the Slovak Karst Biosphere Reserve. Combined with SOC concentration data obtained from 150 sampling points, it was possible to assess landscape-level SOC stock contained in the mosaic of local Rendzic Leptosols, Chromi-Rendzic Leptosols, and Chromic Cambisols. Our estimates confirm that calcareous soils have the capacity to maintain high SOC stocks even under centuries-long agroforestry management that has resulted in a temporally very dynamic mixture of forestland, pastureland, and cropland. When averaged over the entire area, SOC stock reached ca 200 Mg ha^{-1} within 0–60 cm layer, but its spatial variability was at the same time also exceptionally high with 43% coefficient of variation. Therefore, it would have hardly been possible to achieve the study goals by the quantitative pits method or similar. Despite the variability of stoniness, ANCOVA revealed the effect of land use on the SOC concentration. While forestland and pastureland featured practically identical SOC concentrations of about 0.040 g g^{-1}, the concentration in cropland was 0.026 g g^{-1}. The chronosequence analysis indicates that the crop production has been reducing SOC stock at a comparatively low rate of ca 0.002 year^{-1} and that the grazing regime has restored SOC stock back to the forestland level. It appears that favorable SOC retention properties of the calcareous soils provide a broad maneuvering space for land managers in order to use temperate zone agroforestry approaches for offsetting SOC depletion or even achieving SOC accumulation.

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