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Soil organic carbon along an altitudinal gradient in the Despeñaperros nature reserve, Southern Spain

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Abstract

Soil organic carbon (SOC) is extremely important in the global carbon (C) cycle as C sequestration in non-disturbed soil ecosystems can be a sink of C and mitigate greenhouse gas driven climate change. Soil organic carbon changes in space and time are relevant to understand the soil system and its role in the C cycle, and this is why the influence of topographic position on SOC should be studied. Seven topographic positions (toposequence) were analyzed along an altitudinal gradient between 607 and 1168 m.a.s.l. in the Despeñaperros nature reserve (Natural Park). At each study site, soil control sections (25 cm intervals) were sampled. The studied soils are mineral soils with > 3 % organic carbon content. The main characteristic of the studied soils is SOC reduction with depth; these results were related to the gravel content and to the bulk density. The SOC on the surface was highly variable along the altitudinal gradient ranging between 27.3 and 39.9 g kg^{-1}. The SOC stock (SOCS) in the studied area was influenced by the altitude, varying between 53.8 and 158.0 Mg ha^{-1}. Therefore, the altitude factor must be considered in the SOCS estimation at local-regional scale.

1 Introduction

Soils are an important carbon (C) reservoir (Barua and Haque, 2013; Yan-Gui et al., 2013). In fact, the primary terrestrial pool of organic carbon (OC) is soil, which accounts for more than 71 % of the Earth's terrestrial OC pool (Lal, 2010). In addition, soils have the ability to store C for a long time (over the last 5000 years) (Brevik and Homburg, 2004). Soils play a crucial role in the overall C cycle, and small changes in the soil organic carbon stock (SOCS) could significantly affect atmospheric carbon dioxide (CO_2) concentrations, and through that global climate change. Within the C cycle, soils can be a source of greenhouse gases through CO_2 and methane (CH_4) emissions, or can be a sink for atmospheric CO_2 through OC sequestration in soil organic matter (SOM) (Breuning-Madsen et al., 2009; Brevik, 2012).

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Climate, soil use and soil management affect C variability, particularly in soils in dry Mediterranean climates characterized by low OC content, weak structure and readily degradable soils (Hernanz et al., 2002). In temperate climates, recent studies show differences in C sequestration in soils depending on the use and management (Zinn et al., 2007), climate and mineralogical composition (Wang et al., 2010), texture, slope and altitude (Hontoria et al., 2004), and tillage intensity and no-till duration (Umakant et al., 2010). Soil conservation strategies are being seen as a strategy to increase the SOM content (Barbera et al., 2012; Batjes et al., 2014; Jaiarree et al., 2014; Srinivasarao et al., 2014; Fialho and Zinn, 2014).

Several studies have been carried out to estimate differences in soil organic carbon (SOC) in relation to soil properties, land uses and climate (Eshetu et al., 2004; Lemenih and Itanna, 2004). Although the impact of topographic position on soil properties is widely recognized (Venterea et al., 2003; Fu et al., 2004; Brevik, 2013), relatively few studies have been conducted to examine the role of topographic position on SOC content (Ruiz-Sinoga et al., 2012).

The spatial variation of soil properties is significantly influenced by some environmental factors such as topographic aspect that induced microclimate differences, topographic (landscape) positions, parent materials, and vegetation communities (Johnson et al., 2000; Ollinger et al., 2002; Brevik, 2013). Ovales and Collins (1986) evaluated soil variability due to pedogenic processes across landscapes in contrasting climatic environments and concluded that topographic position and variations in soil properties were significantly related. McKenzie and Austin (1993) and Gessler et al. (2000) found that variations of some soil properties could be related to the slope steepness, length, curvature and the relative location within a toposequence. In both cases, the hillslope sequence could be used to understand soil property variations in order to establish relationships between specific topographic positions and soil properties. Asadi et al. (2012) found that the integrated effect of topography and land use determined soil properties. Topography is a relevant factor as it controls soil erosion processes and through that the redistribution of soil particles and organic matter (OM) (Cerdà and García Fayos, 1997; Ziadat and Taimeh, 2013).

Within the body of research covering the spatial distribution of soil properties (Fernández-Calviño et al., 2013; Haregeweyn et al., 2013; Ozgoz et al., 2013; Wang and Shao, 2013) the study of the topographic factor has been included. Over time, many researchers have quantified the relationships between topographic parameters and soil properties such as OM and physical properties such as particle size distribution, bulk density and depth to specific horizon boundaries (McKenzie and Austin, 1993; Gessler et al., 1995; Gessler et al., 2000; Pachepsky et al., 2001; Ziadat, 2005). Organic matter content has been negatively correlated with the topographic gradient (Ruhe and Walker, 1968), and SOC was correlated with slope gradient (Nizeyimana and Bicki, 1992). However, quantitative relationships between soil topography and soil physical-chemical properties are not well established for a wide range of environments (Hattar et al., 2010).

Research along altitudinal gradients has shed light on the effects of climate on soil properties. Ruiz-Sinoga et al. (2012) found a strong relationship between SOM and altitude, which was due to reduced SOM decomposition rates with lower temperatures. High erosion rates have been found in the driest climates (lowest altitudes) such as in Israel (Cerdà, 1998a, b), which support the idea of high OM losses due to surface wash in the driest (lowest altitude) climates. Similar results were found by Ruiz-Sinoga and Martínez-Murillo (2009) in their study on the hydrological response of soil along a climatological gradient in Andalucía, Spain. Ruiz-Sinoga and Diaz (2010) found that the climatological (altitudinal) factors determined soil degradation rates in the pluviometric gradient they studied in southern Spain.

Within the Despeñaperros nature reserve there is no information about the soil variability, and little data is available related to the control topography exerts on soil properties (Lozano-García and Parras-Alcántara, 2014). Therefore, the aims of this study are: (i) to quantify SOC contents and their vertical distribution in a natural forest area, and (ii) to assess the SOCS differences in soils along an altitudinal gradient and their relationship with soil depth in a Mediterranean natural area.

2 Material and methods

2.1 Study site

The 76.8 km^2 Despeñaperros nature reserve (Natural Park) is one of the best-preserved landscapes in southern Europe. It is located within the Eastern Sierra Morena (North of Jaén in Andalusia – Southeast Spain) between 38°20′ and 38°27′ N, 3°27′ and 3°37′ W. Temperatures are low in winter (−10 °C is the lowest daily average temperature) and high in summer (42 °C is the maximum daily average temperature) and the average temperature is 15 °C. The study area is characterized by warm dry summers and cold moist winters. The average annual rainfall is 800 mm, the climate is temperate semi-arid with continental influence features due to altitude and the moisture regime is dry Mediterranean; high temperatures and long drought periods cause water deficits up to 350 mm annually.

It is a mountainous area, with an altitudinal range of 540 m.a.s.l. in the Despeñaperros River Valley to 1250 m.a.s.l at Malabrigo Mountain. The relief is steep with slopes ranging from 3 to 45 %, and the parent materials are primarily slates and quartzites. According to IUSS Working Group WRB (IUSS-ISRIC-FAO, 2006), the most abundant soils are Phaeozems (PH), Cambisols (CM), Regosols (RG) and Leptosols (LP). Well-preserved Mediterranean woodlands and scrublands occupy the study area and large game habitat is the main land use.

2.2 Soil sampling and analytical methods

A toposequence with seven topographic positions (south position-aspect) were selected in the Despeñaperros nature reserve (Natural Park) (Table 1). At each topographic position, soil samples were collected in a random sample design according to

FAO (2006). Each sampling point was analyzed using soil control sections (SCS) at different depths (S1: 0–25, S2: 25–50 and S3: 50–75 cm). SCS were used for a more uniform comparison between studied soils. Four replications for each sampling point were performed in the laboratory (17 sampling points × 1, 2 or 3 SCS × 4 replications).

The soil samples were air-dried at constant room temperature (25 °C) and sieved (2 mm) to remove coarse soil particles. The analytical methods used in this study are described in Table 2.

Statistical analysis was performed using SPSS Inc. (2004). The physical and chemical soil properties were analyzed statistically for PH, CM, RG and LP soils (by SCS), including the average and standard deviation (SD). The statistical significance of the differences in each variable between each sampling point (SCS) was tested using the Anderson-Darling test at each control section for each soil type. Differences with $p < 0.05$ were considered statistically significant.

3 Results and discussion

3.1 Soil properties

The Despeñaperros nature reserve soils are siliceous due to their parent materials (slate, quartzite and sandstone). The studied soils were classified as Phaeozems, Cambisols, Regosols and Leptosols (IUSS-ISRIC-FAO, 2006) (Table 1). The soils are stony soils, acidic, with low base concentrations, oligotrophic and with slightly unsaturated complex change and located in areas of variable slopes ranging between 5 and 38 %. Phaeozems are the most developed soils in the study area. They are deep, dark, and well humidified with high biological activity and high vegetation density on gentle slopes and shady side foothills. Cambisols are developed and deep soils; however Leptosols are the least developed and shallowest soils.

Phaeozems are the most pedogenically developed soils in the study area. They are found on gentle slopes (< 3 %), usually in shady areas on Ordovician sandstones. The gravel content is variable, ranging between 7 and 31 %. Texturally they are sandy soils at the surface and silty-clay-loam or silty-clay soils at depth, with a horizons sequence A0/A1/AB/Bt/C1. These soils show luvic (lv) characteristics (luvic-Phaeozems (lv-PH)) and are > 1 m in depth with pH along the profile ranging from 6.3 to 5.6 at depth and about 4.3 % OM content (Tables 1 and 3).

Cambisols are less developed soils than luvic-Phaeozems, however, these soils are more developed and deeper than Regosols and Leptosols. They appear in areas of variable slope (3–38 %) and are > 1 m in depth characterized by a cambic horizon (Bw) on Ordovician quartzites (Table 1) with approximately 20 % gravel content. At the surface they are sandy soils (< 60 % sand content) with high clay content in the Bw horizon and increasing clay content with depth (Table 3). The horizon sequences were A0/A1/AB/BW/BC/C1 or A0/A1/AB/BW. These soils are characterized by low OM content at depth. Gallardo et al. (2000) showed that the low OM content could be explained by the semiarid Mediterranean conditions. In addition, Parras-Alcántara et al. (2013a) found there is less OM and fewer mineral aggregates in sandy soils, thus favoring high levels of OM transformation. Because of this, Hontoria et al. (2004) suggested that physical variables determine soil development in the driest areas of Spain to a greater degree than management or climatic variables. The Cambisols topsoil has humic (hu) characteristics, with > 5 % OM content (Table 3) due to plant debris accumulation in the A0 horizon. This OM is poorly structured and partially decomposed, thereby reducing the amount and increasing the OM evolution degree with depth. In this line, Bech et al. (1983) reported that the free OM concentration in the surface horizon was higher than 90 %, while humic and fulvic acid concentrations were less than 2 % in soils with *Quercus ilex* spp. ballota vegetation. Free OM was reduced and humidification increased up to 30 % in deeper layers.

Regosols can be found in steeply sloping areas (> 8 %) characterized by high water erosion and subject to rejuvenation processes. We found eutric (eu), dystric (dy) and umbric (um) Regosols (Table 1) on sandstone and quartzite parent materials with > 25 % gravel content in surface layers that eventually disappeared with depth. These soils are sandy-loamy in surface layers and silty-clay in deep layers, with different horizon sequences (A0/A1/AB/BC/C1, A0/A1/AC/C1 and A1/AC/C1). Eutric-Regosols are deeper soils (> 80 cm) that are loamy with high gravel content (25.1–32.2 %) at the surface decreasing with deep, acid pH (5.9) and high OM content (6.7 %) at the surface. The dystric-Regosols are stony soils that are shallow (< 40 cm), loamy at the surface and sandy at depth with high gravel content (> 40%) at the surface, acid pH (6.2) and high OM content (7.3 %) in the surface horizon (Table 3). The umbric-Regosols are also stony, they are deep soils (> 70 cm) that are loamy with high gravel content (40 %) in the surface decreasing to 11 % at depth, acid pH (5.6) and high OM content (6.5 %) (Table 3).

Leptosols are the least developed soils of the study area. Lithic (li), mollic (mo) and eutric (eu) Leptosols were identified (Table 1) formed in sandstones, quartzites and slates on variable slopes (1.5–46 %). Horizon sequences A1/AC/C1, A1/AC, and AC/C1 and A1 were found. The gravel content was variable (> 40 % in the topographically elevated areas and decreasing with depth) with high sand content (> 50 %) in the surface layers. One characteristic of these soils is that the clay content increased with depth, reaching up to 30 %. According to Recio et al. (1986), the physical-chemical properties of the soils in the study area are due to lithology, while their low edaphic development is conditioned by the formation age (Porta et al., 2003). According to Nerger et al. (2007) the alteration and pedogenesis processes taking place in these soils usually occur on low slopes. The lithic-Leptosols are the least developed soils at this study site, with thicknesses ranging between 10 and 15 cm in areas of steep slope. In flat areas their low development is due to their extreme youth. These soils are loamy with a high gravel content (> 28 %), acid pH and > 4 % OM content. Mollic-Leptosols are characterized by mollic surface horizons (thick, well-structured, dark, high base saturation and high OM content), on variable slopes (18.5–38.5 %). According to Corral-Fernández et al. (2013) these soils are characterized by organic residue accumulation in the surface horizons; this OM is poorly structured and partially decomposed at the surface with increasing decomposition degree with depth. Umbric-Leptosols are characterized by

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high OM content, are shallow, and either loamy with high stony content (> 20 % gravel content) or sandy (> 55% sand content), have low bulk density conditioned by the OM content, high porosity and acid pH (Table 3).

3.2 Soil organic carbon (SOC) distribution

The soils in the Despeñaperros nature reserve are characterized by > 3 % OC content, making them part of the 45 % of the mineral soils of Europe that have between 2 and 6 % OC content (Rusco et al., 2001). In general, the SOC content decreased with depth at all topographic positions (A, B, C and D positions) (Table 4). However, this property cannot be observed in the lowest topographic positions (E, F and G positions) due to the low edaphic development (umbric-Leptosols, lithic-Leptosols and mollic-Leptosols) as only one SCS exists (S1: 0–25 cm) (Tables 1 and 4).

The soils in this study are characterized by high sand content at the surface (S1) varying between 59.2 and 34.2 % for C and F positions respectively, and reduced sand content with depth in all studied soils (Table 3). Therefore, this high sand content influenced the OM development, giving OM that is poorly structured and partially decomposed and increasing OM development with depth due to sand content reduction and the clay content increase; clay content reaches 45 % in C: S3. In addition, the mineral medium may play an important role in soil humidification processes, so we can explain low SOC concentrations with depth due in part to soil texture, because SOC tends to decrease with depth in virtually all soils whether or not texture changes or regardless of what kind of change in texture occurs. Clays over sands would have a decrease in SOC with depth also, and probably a more marked decrease. In addition, the formation of structural aggregates made up of SOC and the mineral fraction is reduced, thus favoring high OM levels in sandy soils at depth (González and Candás, 2004). Furthermore, Gallardo et al. (2000) argue that the low concentrations of OM at depth can be explained by the climate (Mediterranean semiarid). Similar results have been found by Corral-Fernández et al. (2013), Parras-Alcántara et al. (2014) and Lozano-García and Parras-Alcántara (2013a) in the Pedroches Valley near the study area.

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Another important point to note is that the clay fraction increased with depth in the B and C positions (reaching a clay content of as high as 45 % (C: S3)) and its relation with SOC at S2 (25–50 cm), which was characterized by high SOC values as compared to S3 (B: 2.0/0.6 %; C: 1.8/0.06 %) (Table 4). Burke et al. (1989) and Leifeld et al. (2005) have shown high SOC levels in soils with high clay content indicating clay stabilization mechanisms in the soil. This effect can be observed in the B and C topographic positions, where an increase in clay content was observed at depth as compared to the upper horizons (B: S1-17.2 %/S2-22.1 %; C: S1-16.1 %/S2-35.7 %). This SOC increase may be due to carbon translocation mechanisms (dissolved organic carbon), soil biological activity and/or the root depth effect (Sherstha et al., 2004).

The SOC appears to be concentrated in the first 25 cm (S1) due to OM accumulation, where the mineralization and immobilization C processes should be active. In these mineral soils, the SOC in deeper layers generally follows a non-linear reduction and this relationship may be expressed as an exponential function (Hiederer, 2009). This non-linear distribution with depth were linked to the unequal OM concentrations that were found in the different SCS. In the surface layer (S1), SOC was variable along the toposequence studied ranging between 39.9 and 27.3 g kg^{-1} at the B and F positions, respectively (Table 4). In this regard, it is important to point out that the S1 layer can reach over 60 % of the total SOC (T-SOC) values documented, corresponding to 60, 64.4 and 63 % for the B, C and D positions respectively as compared to the rest of the soil profile (S2 or S2+S3). Batjes (1996) states that for the 0 to 100 cm depth approximately 50 % of SOC appears in the first 30 cm of the soil. Jobbagy and Jackson (2000) showed that 50 % of SOC is concentrated in the first 20 cm in forest soils to 1 m deep. Civeira et al. (2012), who showed that SOC in the upper 30 cm of soils in Argentina is much higher than in the 30–100 cm interval. The data provided by these authors and the results obtained in this study may be comparable because in this study we used a 75 cm depth and the mentioned authors used a 1 m depth. Also, we used SCS with 25 cm increments and they used SCS with 30 and 20 cm increments, therefore, there are not significant differences between our research procedures and

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the procedures used by Batjes (1996), Jobbagy and Jackson (2000) and Civeira et al. (2012) to investigate SOC distribution with depth. Furthermore, Jobbagy and Jackson (2000) indicated that changes in SOC were conditioned by vegetation type (which determines the vertical distribution of roots) and to a lesser extent the effect of climate and clay content. Despite this, climatic conditions can be a determining factor in the SOC concentrations for surface horizons, whereas clay content may be the most important element in deeper horizons. At the regional-global scale SOC increases with precipitation and decreases with temperature (Post et al., 1982).

The T-SOC analysis in the studied area indicated that there were not big differences between T-SOC along the toposequence. In this regard, T-SOC depended on the degree of development of the soil that appeared at each topographical position. The T-SOC was highest at the B (66.5 g kg^{-1}), D (58.1 g kg^{-1}) and C (52.3 g kg^{-1}) positions, corresponding to the Cambisols-Regosols-Leptosols, Regosols, and Phaeozems-Cambisols-Regosols respectively. Leptosols were the soils with the lowest T-SOC with 27.3, 31.9, 32.7 and 38.1 g kg^{-1} at the F, G, E and A topographic positions, respectively. Similarly, it was noted that in deeper soils (B, C and D) > 60 % of SOC occurred in the S1 layer.

In the toposequence studied, there was a variation in precipitation and temperature (the highest topographic positions had more precipitation and lower temperatures compared to lower topographic positions). Total SOC was not affected by climatic variations, but depended on the soil development in each landscape position. In this line, we observed a T-SOC reduction at the lowest topographic positions where the soils were less developed and a T-SOC increase at the highest topographic positions in the more developed soils. In this line, Power and Schlesinger (2002) explained that the topographic position affects T-SOC, because at low temperatures OM decomposition takes place slowly.

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3.3 Soil Organic Carbon Stocks (SOCS)

Soil OCS in the study area showed a reduction with depth in all topographic positions (Table 4). This SOCS reduction along the profile is linked to OM reduction with depth, this reduction in SOCS also depended on the gravel content and the bulk density (Table 3).

When the first SCS (S1: 0–25 cm) was analyzed we observed high SOCS values as high as 91.1 Mg ha^{-1} in the elevated topographic positions (highest value at the B position). The lowest SOCS values were found at the G position (53.8 Mg ha^{-1}), the lowest site in the toposequence. This trend of decreasing SOCS with decreasing elevation is constant except at the A and E positions. This was caused by the soil type, mollic-Leptosols at the A position and umbric-Leptosols at the E position. Both are poorly developed soils with high OM content in the surface horizon).

We observed that at the D and B topographic positions between 53.8 and 58 % of SOCS, respectively, occurred in the S1 SCS. This constituted 63 and 60 % of T-SOC in these topographic positions. This shows that the gravel content and bulk density affects the SOCS in the surface horizons of the toposequence studied, and therefore a SOCS reduction occurs with respect to SOC. In the most developed soil we observed similar SOC and SOCS concentrations (B: 60 %-SOC; 58 %-SOCS) in the S1 layer, conditioned by bulk density and gravel content. In addition, SOCS were reduced at depth conditioned by gravel reduction and bulk density increased. By contrast, Tsui et al. (2013) and Minasny et al. (2006) explained this bulk density decrease with depth by showing that high OM content at the surface was linked to low clay concentrations (Li et al., 2010). In this line, we observed that high SOCS depended on the SOC concentration and the clay fraction, however, the SOC concentration affected the SOCS to a lesser degree so that in S2 (25–50 cm) we found > 10 % of SOCS related to SOC (C position).

By contrast, low SOCS can be found in S3 (50–75 cm) except at the B topographic position (19 Mg ha^{-1}). This situation could be due to the fact that pedological horizons were generally different than the SCS divisions (S1: 0–25 cm; S2: 25–50 and S3: 50–75 cm) (Hiederer, 2009); in other words, the SCS divisions often led to the mixing of two or more soil horizons (depending on thickness horizon) in any given SCS division.

In all studied soils, the clay content increased with depth. This clay content increase is linked to higher values of SOC (B: S2 and C: S2). In this line, we can explain high SOCS concentrations in clayey soils caused by clay stabilization mechanisms on SOC, this effect can observed at the A topographic position which has higher clay content with respect to the B and D positions. However, we can observed a SOCS increase. This is the case at the D and C topographical positions with SOCS values of 52.1 Mg ha^{-1} and 50.1 Mg ha^{-1} respectively in the S2 sampling layer (showing a correlation between S1 and S2), due to carbon translocation processes as dissolved organic carbon, soil fauna and/or the effect of the vegetation rootings in depth (Sherstha et al., 2004).

3.4 Soil organic carbon stocks (SOCS) along the altitudinal gradient

The SOCS results along the toposequence were also studied. In this respect, it is important to point out that the total SOCS (T-SOCS) were influenced by topographical position in the toposequence analyzed. The T-SOCS increased as we ascended in the toposequence in the study area from the lowest elevation position (G: 607 m.a.s.l.) to the highest elevation position (B: 1009 m.a.s.l.), with the exception of the highest topographic position (A: 1168 m.a.s.l.), with a linear regression relationship (Fig. 1) ($y = 0.1034x + 3.5157$; $R^2 = 0.2668$). Similar results were found by Ganuza and Almendros (2003), Leifeld et al. (2005) and Fernández-Romero et al. (2014); each of these studies showed that the T-SOCS increased with altitude. However, Avilés-Hernández et al. (2009) found that T-SOCS decreased with altitude in forest soils in a toposequence in Mexico due to variations in the OM decomposition rate as a result of the different vegetation types found in the different topographic positions; and Lozano-García and Parras-Alcántara (2014) found that T-SOCS decreased with altitude in a traditional Mediterranean olive grove due to erosion. With respect to the A position in this study, the lower T-SOCS (72.9 Mg ha^{-1}) values with respect to the rest of the studied toposequence may be due to soil loss caused by erosion processes in soils with a low level of development. Similar results have been found by Parras-Alcántara et al. (2004) and Durán-Zuazo et al. (2013). Parras-Alcántara et al. (2004) explained their findings as being due to higher values of soil loss due to high erosivity, high erosionability, steep slopes, low vegetation and the lack of conservation practices in the studied area; while Durán-Zuazo et al. (2013) explained this effect by low vegetation densities in the upper parts of mountain areas that can cause high erosion with strong water runoff. Martínez-Mena et al. (2008) have emphasized the effects of erosion on soil C loss, especially in semi-arid conditions. In this context, a low vegetation ratio can accelerate OM decomposition, weakening soil aggregates (Balesdent et al., 2000; Paustian et al., 2000). Cerdà (2000) indicated that this effect (OM decomposition and aggregate destruction) could occur regardless of climatic conditions.

As can be seen in Table 4, the T-SOCS reduction did not occur gradually. In some cases rapid changes were found, while in other situations gradual changes were noted. Abrupt changes in T-SOCS occurred between the B/C and D/E topographic positions, showing T-SOCS differences of 38 $Mg\,ha^{-1}$ and 44 $Mg\,ha^{-1}$ respectively. Gradual changes in T-SOCS occurred between the C/D, E/F and F/G topographic positions with variations of 3 $Mg\,ha^{-1}$, 13 $Mg\,ha^{-1}$ and 6 $Mg\,ha^{-1}$ respectively. Many authors have concluded that the SOCS reduction can be explained by soil physical properties – mainly texture (Corral-Fernández et al., 2013; Parras-Alcántara et al., 2013b). The studied soils are sandy at the surface with higher clay content at depth (soils that have S2 and/or S3 SCS), therefore, OM stabilizing mechanisms are produced, reducing the aggregate formation between SOC and mineral fraction at depth. As a result, the SOCS content is lower with sandy soils (Nieto et al., 2013). González and Candás (2004) and Parras-Alcántara et al. (2013a) obtained similar results, the first in sandy-loamy soils and the second in Mediterranean clayey soils. In addition, low SOC levels are conditioned by the climatic characteristics of southern Europe (Gallardo et al., 2000).

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4 Conclusions

Soils found in the Despeñaperros nature reserve include Phaeozems, Cambisols, Regosols and Leptosols. Phaeozems are the deepest and most developed soils, and Leptosols are the least developed and shallowest soils. These soils are characterized by low OM content with depth due to the semiarid Mediterranean conditions and the high sand content. The studied soils are characterized by organic residue accumulation in the surface horizons.

The SOC content decreased with depth at all topographic positions and the clay fraction increased with depth. The mineral medium played an important role in soil humidification processes. In addition, the SOC in the S2 layers is characterized by high SOC values with respect to the S3 layers indicating clay stabilization mechanisms in the soil. We can explain this increase due to carbon translocation mechanisms (dissolved organic carbon), soil biological activity and/or the root depth effect.

With respect to T-SOC content, there is not a large difference between T-SOC along the toposequence. The T-SOC of these soils depends on the degree of development of the soils found at each topographic position. We can observe a T-SOC reduction at the lowest topographic positions for less developed soils and a T-SOC increase at the highest topographic positions in the more developed soils. SOCS in the study zone show a reduction with depth in all topographic positions. This SOCS reduction along the profile is linked to OM and gravel content reduction and an increase in bulk density with depth. The T-SOCS increased with altitude, due to the higher turnover of organic material (plants) and the lower decomposition rate due to lower temperatures.

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Table 1. Soil groups of the study area at each of the seven topographic positions with properties. The key refers to the reference soil groups of the IUSS Working Group WRB (2006) with lists of qualifiers.

Topographic position	m.a.s.l.[a]	Slope %	Parent material	Vegetation series	Soil groups	Qualifiers	n[b]
A	1168	15.3	Quartzite – Sandstone	Maritime pine (*Pinus pinaster*) Holm oak (*Quercus ilex*) Gum rockrose (*Cistus ladanifer*)	Leptosols – LP	Mollic – mo	2
B	1009	16.5	Quartzite – Sandstone	Holm oak (*Quercus ilex*) Cork oak (*Quercus suber*) Strawberry tree (*Arbutus unedo*) Gum rockrose (*Cistus ladanifer*)	Regosols – RG Leptosols – LP Cambisols – CM	Eutric – eu Mollic – mo Humic – hu	3
C	945	20.8	Quartzite – Sandstone	Stone pine (*Pinus pinea)* Mastic (*Pistacia lentiscus*)	Cambisols – CM Regosols – RG Phaeozems – PH	Humic – hu Dystric – dy Luvic – lv	3
D	865	5.5	Quartzite	Portuguese oak (*Quercus faginea*) Strawberry tree (*Arbutus unedo*) Gum rockrose (*Cistus ladanifer*)	Regosols – RG	Umbric – um	2
E	778	10.7	Quartzite – Slates	Holm oak (*Quercus ilex*) Strawberry tree (*Arbutus unedo*) Gum rockrose (*Cistus ladanifer*)	Leptosols – LP	Umbric – um	3
F	695	12.0	Quartzite	Cork oak (*Quercus suber*) Holm oak (*Quercus ilex*) Strawberry tree (*Arbutus unedo*) Gum rockrose (*Cistus ladanifer*)	Leptosols – LP	Litic – li	2
G	607	18.5	Slates	Holm oak (*Quercus ilex*) Mastic (*Pistacia lentiscus*)	Leptosols – LP	Mollic – mo	2

[a] Metres above sea level; [b] sample size.

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Table 2. Methods used in field measurements, laboratory analysis and to make calculations from study data.

Parameters	Method
Field measurements	
Bulk density (Mg m^{-3})	Cylindrical core sampler[a] (Blake and Hartge, 1986)
Laboratory analysis	
Particle size distribution	Robinson pipette method (USDA, 2004)[b]
pH – H_2O	Volumetric with Bernard calcimeter (Duchaufour, 1975)
Organic C (%)	Walkley and Black method (Nelson and Sommers, 1982)
Parameters calculated from study data	
SOC stock (Mg ha^{-1})	(SOC concentration × BD × d × $(1 - \delta_{2\,mm}\%)$ × 0.1)[c] (IPCC, 2003)
Total SOC stock (Mg ha^{-1})	$\Sigma_{horizon}$ SOC $Stock_{horizon}$ (IPCC, 2003)

[a] 3 cm diameter, 10 cm length and 70.65 cm^3 volume.

[b] Prior to determining the particle size distribution, samples were treated with H_2O_2 (6 %) to remove organic matter (OM). Particles larger than 2 mm were determined by wet sieving and smaller particles were classified according to USDA standards (2004).

[c] Where SOC is the organic carbon content (g Kg^{-1}), d the thickness of the soil layer (cm), δ2 mm is the fractional percentage (%) of soil mineral particles >2 mm in size in the soil, and BD the soil bulk density (Mg m^{-3}).

Table 3. Properties of the soils evaluated (average ±SD*) in the Despeñaperros Nature Reserve.

Topographic position	m.a.s.l. m	SCS	Depth cm	Gravel %	Sand %	Silt %	Clay %	B.D. $Mg\,m^{-3}$	O.M. $g\,kg^{-1}$	pH H_2O
A	1168	S1	0–25	33.1 ± 13.8 aA	56.5 ± 1.1 aA	22.3 ± 3.0 aA	21.2 ± 4.1 aA	1.1 ± 0.19 aA	64.5 ± 8.9 aA	6.3 ± 0.7 aA
		S2	25–50	7.0 ± 3.1 bA	39.3 ± 0.81 bA	30.7 ± 4.2 aA	30.0 ± 6.1 aA	1.5 ± 0.21 bA	0.99 ± 0.21 bA	5.3 ± 0.5 bA
B	1009	S1	0–25	17.0 ± 10.0 aB	52.9 ± 29.8 aA	29.9 ± 30.6 aA	17.2 ± 5.3 aA	1.1 ± 0.10 aA	68.6 ± 5.2 aA	5.9 ± 0.4 aA
		S2	25–50	27.1 ± 6.4 bB	58.7 ± 20.1 aB	19.1 ± 12.2 bB	22.1 ± 8.0 aB	1.3 ± 0.12 aB	35.3 ± 3.4 bB	5.6 ± 0.7 aA
		S3	50–75	14.3 ± 16.9 aA	41.6 ± 18.1 bA	25.7 ± 15.2 aA	32.6 ± 2.9 bA	1.5 ± 0.12 bA	10.5 ± 2.8 cA	5.7 ± 0.5 aA
C	945	S1	0–25	34.0 ± 5.5 aA	59.2 ± 7.2 aA	24.7 ± 3.1 aA	16.1 ± 6.2 aA	1.2 ± 0.10 aA	58.0 ± 9.5 aA	5.9 ± 0.8 aA
		S2	25–50	14.4 ± 7.2 bC	36.1 ± 12.2 bA	28.2 ± 2.5 aA	35.7 ± 14.1 bA	1.3 ± 0.06 aB	30.9 ± 6.3 bB	5.5 ± 0.4 aA
		S3	50–75	14.9 ± 11.9 bA	24.4 ± 15.9 cB	30.4 ± 9.8 aA	45.2 ± 16.2 cB	1.5 ± 0.05 aA	0.99 ± 0.12 cB	5.2 ± 0.6 aA
D	865	S1	0–25	39.9 ± 6.2 aA	47.6 ± 19.3 aB	38.1 ± 7.5 aB	14.3 ± 2.1 aA	1.1 ± 0.09 aA	62.9 ± 10.4 aA	5.6 ± 1.0 aA
		S2	25–50	24.0 ± 4.5 bB	46.6 ± 18.2 aC	36.2 ± 7.9 aA	17.2 ± 5.4 aB	1.3 ± 0.10 aB	35.9 ± 7.6 bB	5.7 ± 0.8 aA
		S3	50–75	11.9 ± 10.2 cA	30.9 ± 11.1 bB	47.1 ± 5.4 bB	22.0 ± 6.8 aC	1.5 ± 0.13 bA	1.0 ± 0.30 cB	4.5 ± 0.4 bB
E	778	S1	0–25	25.5 ± 6.8 aC	52.2 ± 7.2 aA	30.2 ± 5.1 aA	17.6 ± 2.4 aA	1.2 ± 0.13 aA	56.3 ± 8.9 aA	5.7 ± 0.7 aA
F	695	S1	0–25	28.2 ± 7.4 aC	34.2 ± 5.3 aC	41.0 ± 9.8 aB	24.8 ± 2.8 aA	1.2 ± 0.14 aA	46.9 ± 7.4 aB	6.3 ± 0.5 aA
G	607	S1	0–25	42.9 ± 19.3 aD	54.9 ± 4.1 aA	27.7 ± 2.5 aA	17.3 ± 6.6 aA	1.3 ± 0.13 aB	54.9 ± 9.2 aB	6.2 ± 0.7 aA

m.a.s.l.: metres above sea level;
SCS: soil control section; BD: bulk density;
O.M.: organic matter.

* Standard deviation.
Numbers followed by different lower case letters within the same column have significant differences ($P < 0.05$) at different depths, considering the same topographic position.
Numbers followed by different capital letters within the same column have significant differences ($P < 0.05$) considering the same SCS at different topographic position.

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Table 4. Soil organic carbon (SOC) content and soil organic carbon stock (SOCS) (average ±SD*) in the Despeñaperros Nature Reserve.

Topographic position	m.a.s.l. m	SCS	SOC $g\,kg^{-1}$	T-SOC $g\,kg^{-1}$	SOCS $Mg\,ha^{-1}$	T-SOCS $Mg\,ha^{-1}$
A	1168	S1 S2	37.5 ± 16.8 aA 0.58 ± 0.09 bA	38.1 ± 8.4 A	70.8 ± 33.5 aA 2.1 ± 0.57 bA	72.9 ± 17.0 A
B	1009	S1 S2 S3	39.9 ± 10.3 aA 20.5 ± 6.4 bB 6.1 ± 7.8 cA	66.6 ± 8.2 B	91.1 ± 13.2 aB 49.8 ± 14.9 bB 19.1 ± 19.2 cA	158.0 ± 15.8 B
C	945	S1 S2 S3	33.7 ± 8.6 aA 18.0 ± 9.1 bB 0.58 ± 0.09 cB	52.3 ± 5.9 C	67.4 ± 9.7 aA 50.1 ± 22.4 bB 1.8 ± 0.26 cB	119.3 ± 10.9 C
D	865	S1 S2 S3	36.6 ± 7.9 aA 20.9 ± 9.0 bB 0.57 ± 0.09 cB	58.1 ± 5.7 C	62.1 ± 8.9 aA 52.1 ± 16.7 bB 1.9 ± 0.30 cB	116.1 ± 8.6 C
E	778	S1	32.7 ± 13.2 aA	32.7 ± 13.2 A	72.6 ± 25.0 aA	72.6 ± 0.65 A
F	695	S1	27.3 ± 15.1 aB	27.3 ± 15.1 A	59.3 ± 27.3 aC	59.3 ± 27.3 A
G	607	S1	31.9 ± 13.1 aB	31.9 ± 13.1 A	53.8 ± 18.3 aC	53.8 ± 18.3 A

m.a.s.l.: metres above sea level; SCS: soil control section;
SOC: soil organic carbon; T-SOC: total SOC;
SOCS: soil organic carbon stock; T-SOCS: total SOCS.
* Standard deviation.
Numbers followed by different lower case letters within the same column have significant differences ($P < 0.05$) at different depths, considering the same topographic position.
Numbers followed by different capital letters within the same column have significant differences ($P < 0.05$) considering the same SCS at different topographic position.

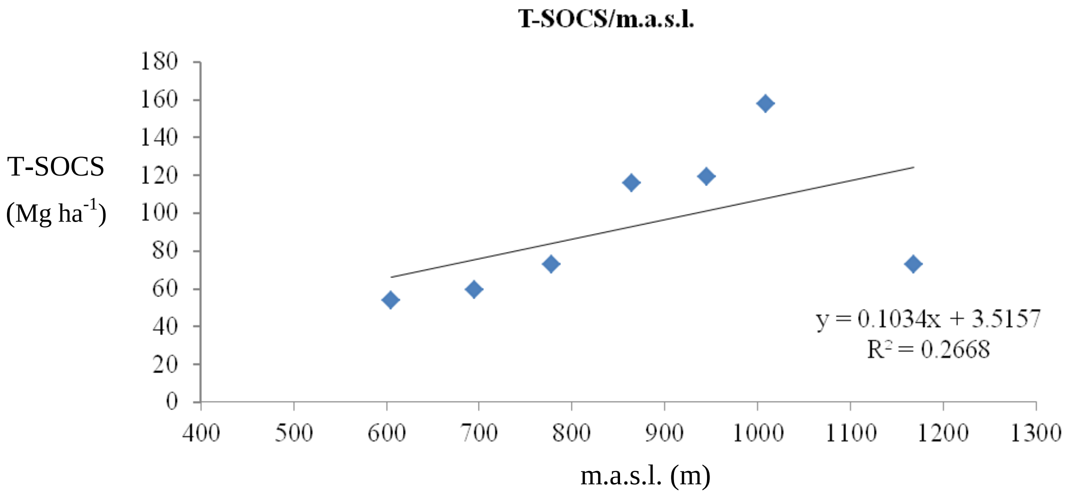

Figure 1. Linear regresion model for T-SOCS versus altitudinal gradient. T-SOCS: total soil organic carbon stock; m.a.s.l.: metres above sea level.

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