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Soil organic carbon and mineral interactions on climatically different hillslopes

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ABSTRACT

Climate and topography have been widely recognized but studied separately as important factors controlling soil organic carbon (SOC) dynamics. Subsequently, the significance of their interplay in determining SOC storages and their pools is not well understood. Here we examined SOC storages and SOC-mineral interactions along two hillslope transects in differing climate zones (MAP = 549 mm in semi-arid eucalyptus savannah vs. 816 mm in temperate eucalyptus forest) in southeastern Australia. On eroding slopes, SOC inventories were twice as large at the wetter site (4.5 ± 0.6 vs. 2.3 ± 0.9 kg m^{-2}), whereas depositional soils had similar SOC inventories at both locations (7.5 ± 2.0 vs.7.0 ± 2.2 kg m^{-2}). On eroding slopes, carbon concentrations of the mineral-associated SOC fraction (< 250 μm and > 2.0 g cm^{-3}) increased by ~50% with increasing rainfall, which was also positively correlated to abundances of clay minerals and pedogenic iron oxides. Within individual hillslopes, carbon concentrations of the mineral-associated SOC fraction doubled from eroding to depositional soils at the drier site, but no topographic trend was observed at the wetter site. The effects of topography on SOC inventories and mineral-associated SOC were more strongly expressed under the drier climate, where vegetation was sparser and soil erosion involved mineral grain size sorting. Our results demonstrate that SOC pools and their interactions with minerals are dependent on topographic locations, emphasizing the need to include geomorphic data when assessing climatic controls of SOC.

1. Introduction

The earth's soil covered landscapes are geomorphically dynamic and under persistent actions of erosion and deposition. Over the past decades, researchers have made substantial progress in understanding how these geomorphic processes affect soil carbon (C) storage and their biogeochemical characteristics (see the review in Doetterl et al., 2016). It has been estimated that global soil erosion (mainly splash and rain-impacted flow) redistributes 1.6 Gt of soil organic carbon (SOC) annually (Müller-Nedebock and Chaplot, 2015), with most of this eroded carbon being re-deposited locally (Chaplot and Poesen, 2012). However, little is known about how minerals in these sediments interact with and influence the redistribution of eroded SOC. While the fate and extent of stabilization of eroded SOC over the course of sediment transport varies widely (Kirkels et al., 2014; McCorkle et al., 2016), some studies have revealed that deposition of eroded soils may accelerate mineral protection of organic carbon (OC) (Doetterl et al., 2015a; Wang et al., 2014).

Underlying the erosional control of SOC is climate, which exerts a strong first-order influence on biological and chemical activity (Amundson, 2001; Giardina et al., 2014). Climatic and geomorphic conditions collaboratively affect plant carbon inputs into soils and microbial decomposition (Luizão et al., 2004; Luo et al., 2017; Yoo et al., 2006), and the same can be stated for the processes of weathering and erosion (Dixon et al., 2016; Riebe et al., 2004; Zapata-Rios et al., 2015) which control soil texture (Attal et al., 2015; Riebe et al., 2015) and thicknesses (Follain et al., 2006; Scarpone et al., 2016). Rates and mechanisms of physical soil erosion differ not only with topographic positions but also with climate (Montgomery and Dietrich, 2002), and substantial particle sorting occurs under certain erosion mechanisms (Doetterl et al., 2016; Heimsath et al., 2002; Shi et al., 2012). For example, water-mediated soil erosion can breakdown macro-aggregates and preferentially transport and redistribute OC-enriched particles from upslope positions, consequently increasing SOC contents in depositional

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areas (Kinnell, 2001; Müller-Nedebock et al., 2016; Oakes et al., 2012). However, it is also known that the transport and loss of particulate organic carbon (POC) by water erosion is significantly affected by climate conditions (Müller-Nedebock and Chaplot, 2015). These examples highlight a critical knowledge gap, wherein climate may impart its signal on the landscape redistribution of SOC erosion by modulating the dominant erosion mechanisms.

The significance of combining these contrasting biological vs. geomorphic perspectives of climate as a primary controller of SOC dynamics is further emphasized by the growing appreciation of mineralogical controls on carbon cycling in soils (Kleber et al., 2015; Lawrence et al., 2015) and recent evidence that SOC is subject to interactive effects between climate and geochemical factors (e.g. soil mineralogy) rather than climatic factors alone (Doetterl et al., 2015b; Kramer and Chadwick, 2016). Contents of clay and iron (Fe) oxides in soils, which often affect soil carbon turnover (Doetterl et al., 2015a; Lawrence et al., 2015; Masiello et al., 2004), are affected by regional climate (Dahlgren et al., 1997; Long et al., 2016) and vary with topographic positions at landscape scales (Berhe et al., 2012; Doetterl et al., 2015a). Lastly, topography can be associated with drastic differences in micro-climate, which may affect soil properties and plant biomass (Luizão et al., 2004; Yoo et al., 2006) as well as organic matter stabilization (Berhe et al., 2012; Chaplot et al., 2001).

How do these interplays between climate and topography affect hillslope soil organic carbon and its associations with reactive minerals? This study's objective is to address this question. Our literature review above reveals significant progresses in understanding landscape redistribution of SOC via erosion and deposition (Doetterl et al., 2016), but it is yet to be fully appreciated that such SOC redistribution is also climate-dependent (Berhe et al., 2012; Doetterl et al., 2015a; Müller-Nedebock et al., 2016). Our understanding of coupling the soil carbon cycle and hillslope processes is in need of being tested across a range of climates, as such efforts will help advance modeling of carbon erosion and landscape-scale soil genesis (Vanwalleghem et al., 2013; Wilken et al., 2017).

In this study, patterns of SOC and the interactions between soil carbon and minerals across two hillslopes under modestly differing climates (MAP = 549 mm in semi-arid eucalyptus savannah vs. 816 mm in temperate eucalyptus forest) were determined. The two sites, which are separated by the great escarpment of southeast Australia, have previously been the focus of various geomorphic and geochemical studies (e.g. Wackett et al., 2018; Burke et al., 2009; Green et al., 2006; Heimsath et al., 2006, 2002; Yoo et al., 2007). Each hillslope spans convex (eroding) to convergent (depositional) landforms, and they share similar soil forming factors (parent material, vegetation, soil age) with the exception of climate. Taking advantage of this climate gradient, our study seeks to (i) identify and assess the significance of climatic and topographic factors that collaboratively affect the size and characteristics of SOC pools and (ii) gain insights into how the formation of the mineral-associated organic carbon pool is attendant to these environmental conditions.

2. Material and methods

2.1. Study site and field sampling

Our study sites, Frog's Hollow (FH) and Nunnock River (NR), are situated above (FH) and below (NR) the Great Escarpment of SE Australia, in a region where small differences in elevation and distance from the coast invoke substantial changes in vegetation from temperate forests to sparsely wooded semi-arid grasslands (Fig. 1). Rolling topography composed of erosional and depositional hillslope segments characterizes the typical landscape of the studied area. The escarpment formed as a result of rifting between Australia and the Lord Howe Rise during the mid-Cretaceous (50–100 Ma; Hayes and Ringis, 1973). Following its formation, the coastward-facing escarpment propagated quickly to near its present location between 35 and 50 Ma and has been roughly stable since (Cockburn et al., 2000; Matmon et al., 2002).

Frog's Hollow (FH), the drier and cooler site above the escarpment, is located ~75 km inland from the coast at an elevation of 930 m. Climate records from Cooma Visitor Center, 37 km south of FH, show 549 mm yr^{-1} of mean annual precipitation (MAP) (ranges from 292 to 843 mm yr^{-1} over the past 45 years). Precipitation occurs throughout the year, but winter (June–August) rainfall is roughly twice that of other months. Mean annual temperature (MAT) is 19.4 °C (minimum of 11.4 °C in July and maximum of 27 °C in January) (Australian Bureau of Meteorology, 2017; http://www.bom.gov.au/climate/data). The warmer and wetter Nunnock River (NR) site is situated in the coastal lowlands directly below the escarpment at an elevation of ~400 m. MAP at Bemboka (7 km east of NR) is 816 mm yr^{-1} (annual low of 307 mm, high of 1678 mm recorded since 1890). MAT at the town of Bega, 30 km east of NR, has an annual maximum of 22 °C (Bureau of Meteorology, Australian Government, Climate averages for Australia, 2017, available at http://www.bom.gov.au/climate/averages/). FH likely experienced periglacial effects from the nearby Snowy Mountains during the Pleistocene, while vegetation changes at NR have been minimal since the last glaciation (Galloway, 1965).

Other than the differing climate, the two sites share similar soil forming state factors (Heimsath et al., 2006; Jenny, 1941). Due to the more arid conditions, FH is covered with a eucalyptus savannah (Fig. 1b), while NR is continuously covered with stringy bark eucalyptus trees. Due to their proximity, the two sites can be viewed as having the same potential flora and fauna, but the existing mix of biota is selected for by local climate (Jenny, 1958). Both sites are underlain by geochemically similar granodiorite. Animal burrowing and tree throw have resulted in a well-defined slope-dependent colluvial flux and soil thickness relationship at NR (Heimsath et al., 2006, 2002), with little to no bedrock exposure. The FH site, due to its sparser vegetation cover, is subject to periodic overland flow in addition to diffusive sediment transport (Heimsath et al., 2001; Wackett et al., 2018), and exposed bedrock tors dot the landscape (Fig. 1b).

Soil production rates were quantified using measurements of the cosmogenic radionuclides ^{10}Be and ^{26}Al and were found to range between 11 and 50 mm Ky^{-1} at FH (Burke et al., 2009; Heimsath et al., 2006) and 15 to 57 mm Ky^{-1} at NR. The rates decreased exponentially with increasing soil thickness (Heimsath et al., 2000). Catchment-averaged erosion rates from ^{10}Be analysis of stream sediments ranged from 13 to 20 mm Ky^{-1} at NR (Heimsath et al., 2000) and 3.6 to 49.4 mm Ky^{-1} at FH (Heimsath et al., 2001). Considering that the soil production and catchment-averaged erosion rates are measured over different spatial scales, it is difficult to assess which of the two sites is eroding faster. Still, these erosion and soil production rates are on the lower side of globally compiled erosion rates that are highly variable (101.6 ± 157.2 mm Ky^{-1}, where the error indicates standard deviation, $n = 663$) (Montgomery, 2007). Both FH and NR soils are classified as Leptic Regosols (IUSS Working Group, 2014).

In total, 23 soil pits with the approximate size of 1 × 1-meter ground surface area were excavated along the hillslope transects: 12 soil pits at FH and 11 pits at NR (Fig. 1b and c). Soil pit locations were selected to include both convex eroding and convergent depositional areas. Depositional hollows at both sites are unchanneled with no indication of gully erosion. Excavations were made to depths of 20–30 cm below the colluvial–saprolite boundary to determine soil thicknesses, and about 1 kg of soil samples were collected by horizon. Saprolite samples were also collected just below the soil-saprolite boundary. Mass percentages of the fine-earth (< 2 mm) and coarse rock fragments (> 2 mm) were determined using a 2 mm sieve in the laboratory. Bulk density cores (5 cm in diameter and 15 cm long) were collected horizontally from soil horizons in excavated soil pits. Collected bulk density cores were oven-dried at 110 °C for a day and then weighed.