Soil burial contributes to deep soil organic carbon storage

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ABSTRACT

Understanding the source of soil organic carbon (SOC) in deep soil horizons and the processes influencing its turnover is critical for predicting the response of this large reservoir of terrestrial C to environmental change and potential feedbacks to climate. Here, we propose that soil burial is a globally important but greatly underestimated process contributing to the delivery and long-term persistence of substantial SOC stocks to depths beyond those considered in most soil C inventories. We draw from examples in the paleosol and geomorphology literature to identify the effects of soil burial by volcanic, aeolian, alluvial, colluvial, glacial, and anthropogenic depositional processes on soil C storage. We describe how the state factors affecting soil formation affect the persistence and decomposition of SOC in buried soils. Organic horizons and surface mineral soils that become buried under layers of sediment can store C several meters below the earth’s surface for millennia or longer. Buried SOC concentrations can rival those of surface soils, and soils buried under volcanic deposits generally contain higher concentrations of SOC than those under alluvial or non-permafrost loess deposits. The dearth of quantitative research on buried SOC specifically, and on deep C pools in general, makes it difficult to estimate the global importance of burial as a terrestrial C storage mechanism on contemporary time scales. The handful of studies that provide data on soil C stocks in buried horizons and estimate their spatial extent suggest that buried soils can contain significant regional OC reservoirs that are currently ignored in inventories and biogeochemical models. Recent research suggests that these buried SOC stocks may cycle biologically on annual-to-decadal time scales if the processes contributing to their protection from decomposition are altered. We discuss the vulnerability of buried SOC pools to disturbance from climate change and human activities that may reconnect these deep SOC pools with the atmosphere. We also provide recommendations on how burial processes can be incorporated into soil biogeochemical models to more accurately predict dynamics of deep SOC pools under different landscapes and environmental conditions.

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1. Introduction

Soils hold one of the largest terrestrial reservoirs of organic carbon (OC), and while most of this pool cycles on very slow time scales (centuries to millennia), climate change and landscape disturbance can affect the proportion of soil organic carbon (SOC) with the potential for more rapid exchange with the atmosphere (Houghton et al., 2001; Trumbore, 2009). Most of our understanding of the processes affecting C storage and release from soils comes from studies of the top 30 cm of soil, which has been the focus of agricultural studies concerned with nutrient cycling (Richter and Mobley, 2009; Zabowski et al., 2011). The long-held assumption that SOC with old radiocarbon (14C) ages is not responsive to changes in environmental conditions has also contributed to underestimates of the amount and dynamics of C stored in deep soils (cf. Fontaine et al., 2007). Evidence that modern 14C is incorporated into subsurface soils (Koarashi et al., 2012) and that ancient SOC can respond to disturbances on annual to decadal time scales (Paul et al., 2006), in conjunction with reinterpretations of relationships between age and reactivity (Torn et al., 2009), provide new insights into the role of deep SOC in the contemporary global C cycle. A growing number of studies report a dynamic response of SOC pools that are located at and below 1 m depth to changes in land use and land cover (Devine et al., 2011; Harrison et al., 2011; Veldkamp et al., 2003).

Recent investigations of deep SOC have focused on belowground root inputs and the vertical transport of particulate and dissolved organic matter through bioturbation, physical mixing, gravity, and
preferential flowpaths as the main modes of delivery of OC to the subsoil (Chabbi et al., 2009; Lorenz and Lal, 2005; Marín-Spiotta et al., 2011; Rumpel and Kögel-Knabner, 2011; Tonneijck and Jongmans, 2008; Wilkinson et al., 2009). Depositional processes have received considerable attention in the context of long-range soil erosion and sedimentation on land (Berhe and Kleber, 2013; Berhe et al., 2012, 2007; Quinton et al., 2010; Van Oost et al., 2007). The role of soil burial in the sequestration of C photo-synthesized in situ at depositional sites has been largely ignored in discussions of deep SOC dynamics. Burial can disconnect a soil from atmospheric conditions and slow or inhibit the microbial decomposition of native or transported SOC. Buried soil horizons, or surface soils that have been buried through various depositional processes, can store more SOC than would exist at such depths from root inputs and leaching from upper horizons (Gurwick et al., 2008a; Hoffmann et al., 2013; VandenBygaart et al., 2012).

Here, we propose that soil burial processes are important contributors to the accumulation and long-term persistence of considerable quantities of SOC at depths beyond those considered by global soil C inventories. While the USDA Soil Taxonomy places the lower limit of soil at 2 m below the surface (Soil Survey Staff, 2010), we expand our consideration of soil to include material above weathered rock that is biologically active and plays an important role in biogeochemical cycling, sensu Richter and Markewitz (1995). Organic matter makes up the bulk of the soil many meters in depth in peatlands (Tarnocai et al., 2009; Ellery et al., 2012) and in permafrost soils (Bockheim and Hinkel, 2007; Johnson et al., 2011; Schuur et al., 2008). Most recent reviews on deep SOC have focused on the top 1 m of soil, whereas here, we focus on SOC in deeper buried soils to highlight the large stores of potentially biologically active C beyond depths commonly measured. Our review excludes very deep paleosols that may have been subjected to metamorphism and behave more like sedimentary rocks.

We review the contributions of geomorphic, pedogenic, and anthropogenic depositional processes to deep SOC storage and describe how environmental conditions or state factors (sensu Jenny, 1941) influence SOC persistence in buried soils, both during and since burial. Understanding the mechanisms contributing to the stabilization or mobilization of SOC from deep buried soils is important for predicting the response of soil C pools to changes in environmental conditions. Buried soils have been studied historically for information about past environments (Valentine and Dalrymple, 1976; Zech et al. 2013) and can also serve as case studies for understanding both the sensitivity of landscape processes to future environmental change and the mechanisms contributing to soil organic matter stabilization (Chaopricha, 2013; Marín-Spiotta et al., in review) Where specific information on factors affecting soils buried below 1 m is scarce, we cite studies of shallower buried soils. Building upon what is known about the processes contributing to the stabilization of deep SOC in buried soils, we discuss vulnerability of these underestimated C pools to landscape and climatic disturbances that may reconnect potentially ancient OC with the atmosphere. Finally, we propose that burial processes and landscape disturbance be incorporated into soil biogeochemical models to better predict the response of belowground OC pools to global change.

2. Buried soils as deep SOC reservoirs

A buried soil is defined by the National Resources Conservation Service as a soil that is “covered with a surface mantle of new soil material that either is 50 cm or more thick or is 30–50 cm thick and has a thickness that equals at least half the total thickness of the named diagnostic horizons that are preserved in the buried soil” (NRCS, 2013). Soil layers underlying a ploughed epipodem, an anthropogenic surface layer with thickness of 50 cm or greater resulting from long-continued manuring, also are considered to be a buried soil. Much of what is known about buried soils comes from the study of paleosols in the geomorphology and Quaternary paleopedology literature (Jacobs and Sedov, 2012). While the term paleosol is sometimes used to describe any soil that developed under different climatic or environmental conditions than today and was once buried, even when overlying material has eroded and it is now exposed at the surface, we use the term here to refer only to currently buried soils (Johnson, 1998). Depositional processes leading to soil burial are certainly not limited to the distant past. Current environmental changes driven or influenced by human activities also lead to landscape disturbances that can result in the rapid burial of recently fixed organic C on land.

Buried SOC can be highly variable spatially and thus challenging to quantify due to the strong dependence of buried soils on landscape geomorphology and climate history (Johnson et al., 2011; Rosenbloom et al., 2006). Furthermore, most studies of buried soils report point locations and do not quantify SOC stocks within the soil profile or across the landscape. Here, we report on a handful of studies that provided either SOC concentrations or soil bulk density and OC concentration data, which allowed us to calculate stocks given the thickness of the horizons. Only two studies we found reported estimates of spatial coverage for buried soils, which allowed for the quantification of reservoir size.

Buried soil horizons can contain large SOC content, comparable to surface soils. One of the few studies on SOC dynamics in deep buried soil horizons, which also reported soil bulk density values, is that of Basile-Doelsch et al. (2005) about a volcanic paleosol sequence on the island of La Réunion. Extrapolating between depth intervals for which specific data were reported, we estimated that the two deepest buried soil horizons sampled (90–276 cm) contained a total of 512 Mg C/ha, or 72% of the whole soil profile SOC to that depth, including surface horizons. Volcanic paleosol sequences below 1 m and up to 3 m depths on substrates of varying age in Mexico contained between 216 and 565 Mg C/ha, 23–50% of whole soil profile SOC (Peña-Ramírez et al., 2009). These soils are located in the Sierra del Chichinautzin Volcanic Field, an area of ca. 2400 km² covered by volcanic deposits with ages between 50,000 and 1670 years B.P. The high SOC densities of these buried soil horizons (with average values of 2.1 ± 0.4 kg/m²) suggest that burial in volcanic deposits contributes to storage of significant SOC stocks deep below the surface. Organic-rich volcanic paleosols buried 3 m below the surface on the slopes of Mt. Kilimanjaro were estimated to contribute a regionally important C “hotspot” of 82 Mg C (Zech et al., 2014).

Buried soils can amount to significant C reservoirs even with low SOC concentrations due to the vast volumes of soil resulting from deep profiles with compacted bulk densities, as well as extensive geographic distributions, especially where buried soils are representative of regional-scale geomorphic processes, as is common with loess deposition (Miao et al., 2007). In the U.S. central Great Plains, a 1-m thick paleosol buried in loess deposits 6 m below the modern surface could contain a maximum of 2.7 Pg C, assuming homogeneous coverage over the large area across multiple states where the soil has been identified (Marín-Spiotta et al., in review). Even if the loess paleosol covered only one tenth of that area, it would still amount to a considerable stock of 0.27 Pg C. These large SOC stocks are located either below SOC stocks or soils accounted for in global C inventories and models and can become exposed to the modern surface through geologic or human disturbances including landslides, road construction, and mining (Chaopricha, 2013; Karlstrom et al., 2008).
In addition to their role in sequestering C from the atmosphere, buried soils can provide important ecosystem services. Deep buried soil horizons with higher SOC concentrations than overlying loess may reduce groundwater contamination by retaining contaminants and inhibiting their downward flow (Duffy et al., 1997). In riparian wetlands, deep buried SOC can influence denitrification potential and thus regulate nitrogen losses (Hill and Cardaci, 2004).

3. Soil burial processes contributing to deep SOC

Storage of SOC at depth results from multiple pedogenic, biotic, and geomorphic factors that contribute to the delivery, accumulation, and persistence of organic matter in the subsurface of soils. Soil burial processes occur in different types of environments and can occur repeatedly over centuries to millennia, resulting in vertical sequences of multiple buried soil horizons (Fig. 1). Soil burial is recognized to be important in the build up of permafrost through successive cryoturbation events (Becher et al., 2013; Bockheim, 2007). SOC in surface organic and mineral soil horizons also can become buried, for example, under lava flows and other volcanic deposits (Basile-Doelsch et al., 2005; Zech, 2006), aeolian loess (Antoine et al., 2013; Mason et al., 2008), flood and alluvial sediments (Blazejewski et al., 2009; Carter et al., 2009), colluvial material landslides and other mass wasting processes (Mayer et al., 2008; VandenBygaart et al., 2012), glacial sediments (Harris et al., 1987; Mahaney et al., 2001), or multiple types of depositional materials (Rawling et al., 2003; Schirrmeister et al., 2013). Human activities can influence the rate and geographic extent of most of these depositional processes as well as contribute to the burial of SOC under impervious surfaces in urbanized or constructed landscapes.

Environmental conditions pre- and post-burial and specific to different depositional processes can influence the types of organic matter that persist in deep buried soils. Not all SOC currently in buried horizons was there at the time of burial. Roots and burrowing animals can extend down into deep horizons and penetrate through multiple paleosols in a soil profile (Gocke et al., 2010) so that it may be difficult to isolate residual SOC from new inputs after the burial process. Disturbance events subsequent to burial can also result in the delivery of more recent C into buried soil horizons. For

Fig. 1. Example of different soil burial processes (top panel) and the paleosol sequences they form (bottom panel): a) volcanic burial at Mt. Kilimanjaro, Tanzania (photo from Zech, 2006); b) aeolian burial at the Old Wauneeta Roadcut in southwestern Nebraska, USA (photo modified from Feggestad et al., 2004); and c) alluvial burial at Pressey Park, South Loup River in central Nebraska, USA (photo by Joseph Mason).
Table 1
Examples of soil organic carbon (SOC) stocks in volcanically buried soil horizons below 1 m depth. Numbers (1, 2) indicate buried soils from different profiles or sites reported in the same study. We report mean and standard error (SE) SOC stocks for buried soils in cases when vertical sequences of multiple buried soils were reported by the authors. Total SOC stocks were calculated by summing SOC stocks for all buried soil layers within a soil profile. Burial times/soil ages are based on dates of the burial deposits or buried horizons as reported by the authors. YBP = years before present; m.a.s.l. = meters above sea level.

<table>
<thead>
<tr>
<th>Buried soil</th>
<th>Depth range (cm)</th>
<th>Mean SOC stocks (kg C/m²)</th>
<th>SE SOC stocks (kg C/m²)</th>
<th>Total SOC stocks (Mg C/ha)</th>
<th>Burial time or soil age (y)</th>
<th>Location</th>
<th>Current vegetation</th>
<th>MAT (°C)</th>
<th>MAP (mm)</th>
<th>Altitude (m.a.s.l.)</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>1) Tlaloc P</td>
<td>95–142</td>
<td>1.6</td>
<td>0.5</td>
<td>62.0</td>
<td>6200 ± 85</td>
<td>Mexico</td>
<td>Pine forest</td>
<td>10–14</td>
<td>1000–1200</td>
<td>3100</td>
<td>Peña-Ramírez et al., 2009</td>
</tr>
<tr>
<td>2) Chautzin P</td>
<td>79–270</td>
<td>3.7</td>
<td>1.3</td>
<td>256.6</td>
<td>1835 ± 55</td>
<td>Mexico</td>
<td>Pine forest</td>
<td>10–14</td>
<td>1000–1200</td>
<td>3100</td>
<td>Peña-Ramírez et al., 2009</td>
</tr>
<tr>
<td>3) Pelado P1</td>
<td>82–200</td>
<td>1.9</td>
<td>1.1</td>
<td>77.8</td>
<td>9620 ± 160 to 10,900 ± 280</td>
<td>Mexico</td>
<td>Pine forest</td>
<td>10–14</td>
<td>1000–1200</td>
<td>3100</td>
<td>Peña-Ramírez et al., 2009</td>
</tr>
<tr>
<td>4) Pelado P2</td>
<td>88–180</td>
<td>3.0</td>
<td>1.8</td>
<td>89.7</td>
<td>9620 ± 160 to 10,900 ± 280</td>
<td>Mexico</td>
<td>Pine forest</td>
<td>10–14</td>
<td>1000–1200</td>
<td>3100</td>
<td>Peña-Ramírez et al., 2009</td>
</tr>
<tr>
<td>5) Malacatepetl P</td>
<td>95–250</td>
<td>0.8</td>
<td>0.5</td>
<td>41.5</td>
<td>30,500 ± 1160</td>
<td>Mexico</td>
<td>Pine forest</td>
<td>10–14</td>
<td>1000–1200</td>
<td>3100</td>
<td>Peña-Ramírez et al., 2009</td>
</tr>
<tr>
<td>6) Malacatepetl A</td>
<td>75–293</td>
<td>2.4</td>
<td>0.3</td>
<td>143.8</td>
<td>30,500 ± 1160</td>
<td>Mexico</td>
<td>Fir forest</td>
<td>10–14</td>
<td>1000–1200</td>
<td>3200</td>
<td>Peña-Ramírez et al., 2009</td>
</tr>
</tbody>
</table>

1) 1750 N 100–210 0.4 0.0 48.7 55,000 YBP to early Holocene Tanzania Up to ~ 1800 m a.s.l.: savanna/cultivated land. Above ~ 1800 m: sub-montane and montane forests Max 3000 mm/year between 2000 and 2400 m a.s.l., much less above and below those elevations 1750 Zech et al., 2014

2) 1850 N 100–250 0.5 0.0 53.7 1850

3) 1400 W 100–130 0.6 0.0 18.2

4) 1700 N 100–200 0.8 0.1 40.0

5) 2800 N 100–300 2.3 0.5 411.1

6) 2200 N 100–160 3.9 0.9 117.9

7) 2600 N 100–210 5.7 0.8 627.4

8) 1700 100–160 1.8 0.2 105.7

9) 2350 100–200 3.6 0.6 285.4

10) 1750 100–215 3.2 0.7 319.0

11) 2140 100–215 10.3 4.5 413.4

12) 2450 100–230 4.7 1.6 282.1

13) 2250 100–235 6.6 0.7 463.9

14) 750 100–300 0.4 0.1 28.5

15) 1430 100–300 2.1 0.1 213.8

16) 2750 115–160 8.1 2.1 242.8

17) 2550 115–213 4.1 0.8 122.5

1) Tephras P 90–276 3.2 1.0 511.6 31,700 YBP and earlier La Réunion Forest 13 1700 1720 Basile-Doelsch et al., 2005

1) Sandy loam 560–590 1.2a 0.1 36.1 79 AD Pompeii, Italy Urban 6.7 Vogel and Märker, 2011

2) Sandy loam 640–670 1.0a 0.3 30.2

3) Sandy loam 550–600 0.5a 0.1 27.3

4) Sandy loam 610–670 1.2a 0.2 70.7

5) Silty loam 610–670 2.2a 0.1 130.8

a Estimated using mean soil bulk density (0.6 g/cm³) for volcanic paleosols from Zech et al. (2014) and Basile-Doelsch et al. (2005).
example, in the Rocky Mountains, the accumulation of black C in deep buried soil horizons was attributed to the incorporation of new inputs from recent surface burning that were transported downwards during clay translocation (Leopold et al., 2011).

In the following sections, we examine different burial processes and discuss how differences in the time scales, geographic extent, and properties of the sediment and depositional environment affect the amount and persistence of SOC. We provide generalizations on the relative amounts of SOC buried under different depositional processes and the frequency and spatial occurrence of specific burial processes with the awareness that geographic biases by researchers investigating buried soils could very well contribute to some of these observed differences. We report values for SOC concentrations (mg C/g soil) for different depositional processes in Tables 1–4 as provided by the authors. For volcanic and loess soils (Tables 1 and 2), we also estimated SOC content (kg/m²) using bulk density values for similar buried soils.

3.1. Volcanic deposits

Tephra and lava flows have buried soils around the world in regions with currently or formerly active volcanoes under widely varying climatic and vegetation types. Volcanic events often occur repeatedly over geologic and historic periods of time, resulting in multiple layers of vertical paleosol sequences such as those found on the southern slopes of Mt. Kilimanjaro in Tanzania (Zech et al., 2014) and near the Nevado de Toluca volcano in Central Mexico (Sedov et al., 2003, 2001). The global extent of soil buried through volcanic deposition tends to be concentrated in specific regions under the influence of paleo or active volcanism, unlike other depositional events with much wider geographic footprints.

Volcanic deposition typically occurs very rapidly, burying live vegetation and organic residues in soils under flows of lava, ash, pumice, or debris avalanches. The properties of the volcanic deposits are likely to affect the rate of burial, the depth of the deposits, and the potential transformations of organic matter. During certain combustion conditions, plant and soil organic matter can be converted into condensed aromatic structures known as black C, which can persist in some soil environments for thousands of years (Spokas, 2010). Burial by volcanically-derived dust during cold and arid glacial periods is thought to have contributed to the accumulation of SOC-rich (100 mg/g soil) paleosol sequences up to 3 m deep on the slopes of Mt. Kilimanjaro in Tanzania (Zech et al., 2014).

Soils developing on weathered lava tend to accumulate high amounts of organic C due to the abundance of noncrystalline or amorphous minerals (Basile-Doelsch et al., 2005; Torn et al., 1997). The high sorptive capacity of most volcanic soils and the presence of black C can lead to the long-term persistence of high amounts of SOC from decomposition, making these buried soils highly vulnerable to changes in climate.

3.2. Loess deposits

Aeolian deposits formed from the accumulation of windblown dust tend to exist in currently or formerly arid areas with a dust source, wind, and minimal vegetative cover. Such conditions have developed through the wind erosion of deserts and during deglaciation as glacial meltwater decreased and floodplains dried up, exposing glacial silts and clays to erosional winds. Loess deposition can cover very large geographic areas such as the U.S. Great Plains (Bettis et al., 2003) and plateaus in western China (Lu and Sun, 2000). The resulting soil building process tends to occur gradually, sometimes over thousands of years, as dust accumulates and soil thickness increases over time. In the U.S. Great Plains, grassland Mollisols aggregated throughout the Holocene as loess accrued in multiple events due to changes in climate and vegetative cover, resulting in sequences of A horizons that developed at the end of the Pleistocene and today are buried up to several meters below the modern soil surface (Jacobs and Mason, 2005).

Our review of the literature revealed that many paleosols in loess deposits tend to have low SOC concentrations (Table 2), which in part may be related to the formation of these soils during relatively brief periods of slower loess deposition in environments with low primary production. Episodes of dune mobilization and wind erosion can be associated with periods of aridity and increasing fire frequency, leading to the accumulation of black C in buried soils overlain by loess sediments (Marin-Spiotta et al., in review; Wang et al., 2005). The highest SOC stocks found in paleosols were those in Siberian yedoma loess permafrost (Table 2), which forms when wind-blown sediments bury roots and other organic matter that then become incorporated into deeper permafrost. Loess permafrost can contain consistently large SOC content down 20 or 30 m (Zimov et al., 2006a). Freezing temperatures protect yedoma SOC from decomposition, making these buried soils highly vulnerable to changes in climate.

3.3. Alluvial and colluvial deposits

Sediment deposited in stream terrace fill, alluvial fans, and overbank floodplain deposits can also bury and stabilize SOC. Thick, dark A horizons buried in alluvial deposits were found in U.S. Central Plains stream valleys as deep as 11 m below ground, where they have been buried for approximately 10,000 years (Mandel, 2008). Floodplain sediments can bury peat, as they have in the Mkuze coastal floodplain of South Africa, protecting organic matter from threats to currently exposed peatlands for approximately 6000 years (Ellery et al., 2012). In Venezuela, a sequence of several peat paleosols was found buried under alluvium, and one buried peat layer at a depth of 13 m from an interglacial period 60–70,000 years ago contained approximately 150–750 mg SOC/g soil (Mahaney et al., 2001).

Riparian zone soils store considerable SOC in buried organic-rich soil horizons. In the Pawcatuck River watershed of Rhode Island, 17 of 22 sites studied contained OC-rich material buried in former surface horizons or lenses beneath alluvial deposits at depth below 1 m and up to 3.5 m (Blazejewski et al., 2009). Buried SOC in saturated riparian areas can take thousands of years to decompose due to low oxygen availability (Blazejewski et al., 2009, 2005) but may become vulnerable to decomposition if drained through human or geomorphic alterations to hydrologic flow. In some areas, riparian zone alluvial and outwash areas have high rates of microbial activity, indicating that buried soils in riparian areas can be
Table 2

<table>
<thead>
<tr>
<th>Buried soil</th>
<th>Depth range (cm)</th>
<th>Mean SOC (g cm⁻²)</th>
<th>SE SOC (g cm⁻²)</th>
<th>Current vegetation</th>
<th>MAT (°C)</th>
<th>MAP (mm)</th>
<th>Altitude (m a.s.l.)</th>
<th>Map (N)</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>1) Buried soil 1</td>
<td>1.70–1.90</td>
<td>0.73</td>
<td>0.01</td>
<td>Shortgrass prairie</td>
<td>6.26</td>
<td>431</td>
<td>600</td>
<td>6.5</td>
<td>Feggestad et al., 2008</td>
</tr>
<tr>
<td>2) Buried soil 2</td>
<td>2.30–2.10</td>
<td>1.05</td>
<td>0.01</td>
<td>Shortgrass prairie</td>
<td>5.20</td>
<td>378</td>
<td>800</td>
<td>3.7</td>
<td>Feggestad et al., 2004</td>
</tr>
<tr>
<td>3) Buried soil 3</td>
<td>1.25–1.35</td>
<td>0.9</td>
<td>0.01</td>
<td>Tidal grass prairie</td>
<td>13.92</td>
<td>216</td>
<td>11,000</td>
<td>2.0</td>
<td>Huang et al., 2002, Karlstrom et al., 2008, Zimov et al., 2006a,b</td>
</tr>
</tbody>
</table>

Note: Estimates using soil bulk density (1.21 g cm⁻³) for boos paleosols from (Matthews et al., in review).
### Table 3
Examples of soil organic carbon (SOC) concentrations of alluvially-buried soil horizons below 1 m depth. Numbers (1, 2) indicate buried soils from different profiles or sites reported in the same study. Letters (a, b) indicate multiple buried soil horizons within a vertical profile. Burial times and soil ages are based on dates of the burial deposits or buried horizons as reported by the authors. YBP – years before present; m.a.s.l. – meters above sea level.

<table>
<thead>
<tr>
<th>Buried soil horizon(s)</th>
<th>Depth (m)</th>
<th>SOC (mg C/g soil)</th>
<th>Burial time or soil age</th>
<th>Location</th>
<th>Current vegetation</th>
<th>MAT (°C)</th>
<th>MAP (mm)</th>
<th>Altitude (m.a.s.l.)</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>1a) Ab3</td>
<td>2.3–3.4</td>
<td>4.5–7</td>
<td>Late Holocene</td>
<td>Southern Plains, Oklahoma, U.S.; 1) Carnegie Canyon 2) Lizard Site</td>
<td></td>
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<tr>
<td>1b) ABb3</td>
<td>3.4–3.5</td>
<td>3</td>
<td></td>
<td></td>
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<td></td>
<td></td>
<td></td>
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<tr>
<td>1c) Bkb3</td>
<td>3.5–4.0</td>
<td>1–2</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>1d) Cb3</td>
<td>4.1</td>
<td>0.5</td>
<td></td>
<td></td>
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<td></td>
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<tr>
<td>2a) A2b</td>
<td>1.3–1.8</td>
<td>7–8</td>
<td></td>
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<tr>
<td>2b) Btssb2/Btbb2</td>
<td>1.8–2.5</td>
<td>3–8</td>
<td></td>
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<tr>
<td>2c) BCb2</td>
<td>2.5–3.3</td>
<td>2–3</td>
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<td>Four buried A horizons under coarse overbank deposits in karst area</td>
<td>2–3.3</td>
<td>15–35</td>
<td>Late Holocene</td>
<td>Rio Indio floodplain, Vega Baja, Puerto Rico</td>
<td></td>
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<tr>
<td>1a) 2Bkb1</td>
<td>1.07–1.35</td>
<td>5.13</td>
<td>Late Holocene</td>
<td>Upper Republican River Basin, Central Great Plains, U.S. 1) Nichols Site 2) State Site</td>
<td></td>
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<tr>
<td>1b) 2Bkb2</td>
<td>1.35–1.60</td>
<td>4.99</td>
<td></td>
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<tr>
<td>1c) 2Bkb3</td>
<td>1.60–2.25</td>
<td>7.22</td>
<td></td>
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<tr>
<td>1d) 2Bw</td>
<td>2.25–3.25</td>
<td>4.04</td>
<td></td>
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<td></td>
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<tr>
<td>1e) 2BC</td>
<td>3.25–4.00</td>
<td>1.01</td>
<td></td>
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<tr>
<td>2a) 4Akb1</td>
<td>0.95–1.12</td>
<td>8.82</td>
<td></td>
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<tr>
<td>2b) 4Akb2</td>
<td>1.12–1.41</td>
<td>10.12</td>
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<tr>
<td>2c) 4Akbb</td>
<td>1.41–1.70</td>
<td>7.22</td>
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<tr>
<td>2d) 4ckb1</td>
<td>1.70–1.95</td>
<td>7.22</td>
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<tr>
<td>2e) 4ckb2</td>
<td>1.95–2.15</td>
<td>5.12</td>
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<tr>
<td>2f) 4ckb3</td>
<td>2.15–2.40</td>
<td>4.16</td>
<td></td>
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<tr>
<td>Interbedded sand and mud stream channel sediments (river channel migration)</td>
<td>1.6–2.7</td>
<td>21.9</td>
<td>Near Toronto, Ontario, Canada</td>
<td>Mixed forest, marsh</td>
<td>205°</td>
<td></td>
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<tr>
<td>Buried Mollisols in floodplains</td>
<td>1.0–5.1</td>
<td>5–40</td>
<td>Since 735 B.C.</td>
<td>Plate River system, Southwestern Wisconsin (Hinderman Site 1)</td>
<td>Agriculture</td>
<td></td>
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<tr>
<td>Floodplain soil</td>
<td>1–5.3</td>
<td>1250–2900 cal YBP</td>
<td>Adigrat, Tigray Plateau, Ethiopia &amp; Eritrea Pawcatuck Watershed, Rhode Island, USA</td>
<td>Grassland</td>
<td>619</td>
<td>2493</td>
<td></td>
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<tr>
<td>1) Ab</td>
<td>1.975</td>
<td>83</td>
<td></td>
<td>Riparian forests</td>
<td></td>
<td></td>
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<tr>
<td>2) Ab</td>
<td>1.005</td>
<td>126</td>
<td></td>
<td></td>
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<tr>
<td>3) Ab</td>
<td>2.025</td>
<td>94</td>
<td></td>
<td></td>
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</tbody>
</table>

a Values were approximated from figures.

b Values were obtained from organic matter amounts using the formula OC = OM/2.
Table 4
Examples of soil organic carbon (SOC) concentrations in soil horizons buried below 1 m depth under glacial, erosional, and multiple types of deposits. Letters (a, b) indicate multiple buried soil horizons within a vertical profile. Numbers (1, 2) indicate buried soils from different profiles or sites reported in the same study. Burial times and soil ages are based on dates of the burial deposits or buried horizons as reported by the authors. YBP – years before present; m.a.s.l. – meters above sea level.

<table>
<thead>
<tr>
<th>Burial processes</th>
<th>Buried soil horizon(s)</th>
<th>Depth (m)</th>
<th>SOC (mg C/g soil)</th>
<th>Burial time or soil age</th>
<th>Location</th>
<th>Current vegetation</th>
<th>MAT (°C)</th>
<th>MAP (mm)</th>
<th>Altitude (m.a.s.l.)</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Till, glaciofluvial, and glaciolacustrine deposition</td>
<td>Vertical sequence of sandy peat paleosols without clay transformation or aeolian deposition</td>
<td>8.6–13.9(^a) 0–750(^\prime)</td>
<td>Wisconsinan/Weichselian Glaciation (up to 60,000 YBP)</td>
<td>Northern Venezuelan Andes</td>
<td>Agriculture: potatoes, grazing</td>
<td>10</td>
<td>100</td>
<td>3550</td>
<td>(Mahaney et al., 2001)</td>
<td></td>
</tr>
<tr>
<td>Colluvium from hillslope erosion in intermontane basin</td>
<td>a) Bkb3</td>
<td>1.05–1.35 1.35–1.85</td>
<td>Holocene</td>
<td>Middle Park, Colorado, USA</td>
<td>Big sagebrush, grasses</td>
<td>12.9 to –5.7 (max/min)</td>
<td>255</td>
<td>2350</td>
<td></td>
<td>(Mayer et al., 2008)</td>
</tr>
<tr>
<td>Colluvium (erosional sediment) deposition</td>
<td>Aggraded sandy loam cropland soil below plow line (Orthic Humo-Ferric Podzols): B horizon</td>
<td>1(^a) 10(^\prime)</td>
<td>Since 1955</td>
<td>Grand Falls, New Brunswick, Canada</td>
<td>Agriculture: potatoes, forage</td>
<td></td>
<td></td>
<td></td>
<td>(VandenBygaart et al., 2012)</td>
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<tr>
<td>Loess, fluvial, and volcanic tephra deposition</td>
<td>a) Paleosol 4</td>
<td>1.0(^a) 2.48</td>
<td>3000–7000 YBP</td>
<td>Stampede site, Cypress Hills, Alberta, Canada with thin, weakly developed paleosols</td>
<td>Aspen forest</td>
<td>1238</td>
<td></td>
<td></td>
<td>(Klassen, 2004)</td>
<td></td>
</tr>
<tr>
<td>Loess, fluvial, and colluvial deposition in aeolian cliff-top deposits</td>
<td>a) Cumulic Ab1</td>
<td>1.00–1.47 1.47–1.52</td>
<td>1263 YBP</td>
<td>S. Dakota White River Badlands, USA</td>
<td>Semi-arid mixed grass ecosystem</td>
<td>10.3</td>
<td>400</td>
<td>950</td>
<td>(Rawling et al., 2003)</td>
<td></td>
</tr>
<tr>
<td></td>
<td>b) Ab1 (darker)</td>
<td>1.52–1.62 1.62–1.80</td>
<td>2003–2036 YBP</td>
<td>2355 YBP</td>
<td>4152–4219 YBP</td>
<td>6665–8588 YBP</td>
<td></td>
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<tr>
<td></td>
<td>c) Ab2</td>
<td>1.80–1.90 1.90–2.05</td>
<td>7</td>
<td>6</td>
<td>5</td>
<td>4</td>
<td></td>
<td></td>
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<tr>
<td></td>
<td>d) Ab2</td>
<td>2.05–2.15 2.15–2.70</td>
<td>5</td>
<td>4</td>
<td>3</td>
<td>2</td>
<td></td>
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<td></td>
<td>e) Cb2</td>
<td>2.70–2.80</td>
<td>3</td>
<td>2</td>
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<tr>
<td></td>
<td>f) Cb3</td>
<td>2.80–2.98</td>
<td>2</td>
<td>1</td>
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<tr>
<td></td>
<td>g) Cb3</td>
<td>2.98–3.20</td>
<td>1</td>
<td></td>
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</table>

\(^a\) Values were approximated from figures.
The redistribution of SOC by agricultural activities is significant: during the last 50 years, an estimated 16–21 Pg C of SOC are thought to have been buried on agricultural lands (Van Oost et al., 2007). While much of this erosion leads to shallow burial, repeated events over time can lead successive layers of sediment. In Belgian croplands under continuous cultivation for at least 250 years, up to 2.5 m of soil accumulated in depositional areas on the landscape, contributing to up to 6 times greater SOC stocks than in eroding areas (Doetterl et al., 2012). Landscape destabilization by human removal of vegetation or alteration of river channels and floodplains can also trigger or magnify the effects of other geomorphic processes (Knox, 2001). A recent analysis of OC burial in soil sediments in agricultural landscapes from Central Europe reported mean OC stocks of 3.1 ± 1.0 kg C/m² and 5.0 ± 1.3 kg C/m² for hillslope and floodplain soils, respectively (Hoffmann et al., 2013).

Few studies have quantified SOC storage under urban environments. One calculation of C storage in U.S. urban soils down to 1 m assumed that SOC pools were negligible below that depth, despite recognizing that this method ignored the SOC in buried A horizons assumed that SOC pools were negligible below that depth, despite recognizing that this method ignored the SOC in buried A horizons (Pouyat et al., 2006). A study of the U.K. city of Leicester, which is 15% covered by pavement, buildings, and other impervious capped surfaces, found no significant difference in SOC stocks in capped vs. greenspace soils at equivalent depths down to 1 m and found higher SOC under capped soils than in nearby arable environments (Edmondson et al., 2012). The authors conclude that soils in urban environments may continue to be influenced by vegetation and lose less SOC over time while conventional agricultural practices tend to deplete soils of OC. It is unknown how construction and other soil disturbance events that are common practice in urbanized environments contribute to the long-term preservation of SOC in soils under anthropogenic impervious surfaces.

4. State factors affecting SOC accumulation and stability in buried soils

The preservation of undecomposed wood dated tens of 1000s of years old in deep buried soils (e.g., Orlova and Panychev, 2006) provides supporting evidence that burial can contribute to long-term persistence of OC and is consistent with current perspectives on the importance of environmental conditions in regulating SOC turnover time (Schmidt et al., 2011). The specific processes contributing to slow turnover times and old radiocarbon ages of SOC in deep soils have been the subject of numerous reviews (Rumpel and Kögel-Knabner, 2011; Salomé et al., 2010; Schmidt et al., 2011; Torn et al., 2009; Trumbore, 2009): depressed microbial biomass and activity due to low rates of oxygen diffusivity, low soil moisture and nutrient availability; the microbial recycling of aged C compounds; and aging during the slow, downward migration of C that escapes mineralization from upper soil horizons. The first two mechanisms contribute to the long-term persistence of SOC present in the soil at the time of burial, while the third affects incorporation of OC following burial. Processes that contribute to slow turnover times for SOC in surface soils and prevent its loss from the soil profile, such as the incorporation of OC into soil aggregates and associations with mineral surfaces, cations or metals (Lorenz et al., 2011; Six et al., 2002; Torn et al., 2009; Von Lützow et al., 2006), also apply to buried mineral soils. Here, we focus on how the state factors of soil formation—climate, organisms, relief, parent material, and time, sensu Jenny (1941)—affect the potential of different burial processes to contribute to the accumulation and long-term persistence of deep SOC. The stability of SOC in a buried horizon is directly influenced by the burial process as well as by the properties of both the buried soil and the overlying depositional material.

4.1. Effects of climate on buried SOC

Climate during soil pedogenesis and at the time of burial can have an important influence on the content, composition, and persistence of OC in deep buried horizons. Climate can affect the magnitude and frequency of specific burial processes; for example, extremely wet and dry periods promote alluvial and aeolian deposition, respectively. In the U.S. Central Great Plains, higher accumulations of SOC in a paleosol of a multiple sequence have been attributed to periods of higher primary production at the time of soil formation during the Pleistocene–Holocene transition, while subsequent increasing aridity led to its burial under successive events of loess deposition (Feggestad et al., 2004). Changes in climate that lead to multiple freeze–thaw cycles are known to influence the rate of incorporation of surface SOC into deeper permafrost layers by vertical mixing (Becher et al., 2013; Koven et al., 2009). In buried permafrost loess and peat soils, frozen temperatures currently preserve deep SOC, but thawing due to future atmospheric temperature are predicted to accelerate microbial decomposition and lead to large OC loss (Schuur et al., 2008; Tarnocai et al., 2009; Zimov et al., 2006a, 2006b).

Deep soils in well-drained profiles tend to have a smaller range of temperatures and lower levels of moisture than surface horizons, which commonly leads to low microbial activity (Rumpel and Kögel-Knabner, 2011; Salomé et al., 2010). Low soil moisture inhibited respiration rates of buried soils in arid loess deposits in the U.S. Central Great Plains (Chaopricha, 2013) and of paleosols collected under burial mounds in the dry Russian steppe (Demkina et al., 2010). The response of paleosols buried 1700 and 4000 years to moisture treatments did not differ, but the magnitude of their response was lower than that of modern surface soils (Demkina et al., 2010). In contrast, waterlogging within a 3330-year-old burial mound in Denmark resulted in marked differences in the microbial biomass and chemical composition of SOC in anaerobic and aerobic microsites (Thomsen et al., 2008).

The depth and structure of overlying deposits affect the rate of gas exchange with the atmosphere as well as the transport of water, nutrients, and microorganisms down to a buried horizon. Most paleosols can be influenced by current surface conditions through slow leaching and translocation. Preferential flow-paths can deliver particulate and dissolved OC from surface organic horizons to mineral subsoils in shrink-swell soils (Marín-Spiotta et al., 2011), and aquifer advection can transport young, surface-derived OC to soil horizons as deep as 11 m (Mailloux et al., 2011). Post-burial changes in groundwater levels can transform paleosols, as evidenced by redoximorphic features in Roman paleosols in the ancient ruins of Pompeii (Vogel and Märker, 2011). These changes in redox can accelerate pedogenesis as well as the mobility of SOC (Hall and Silver, 2013). Future changes in precipitation projected with climate change for many regions of the world can alter current hydrologic patterns that can both reconnect or disconnect buried soil horizons from surface biological processes (Holden, 2005).

4.2. Biotic controls on buried SOC

Plants and soil micro- and macroorganisms, such as animals, bacteria, and fungi, affect the quantity of SOC in soils before and after burial by influencing rates of OC inputs, decomposition and mobilization. High rates of primary production during periods of warming and increased precipitation resulted in the buildup of SOC in soils that were subsequently buried by loess during periods of low vegetative cover that destabilized dune sources (Feggestad et al., 2004). High amounts of SOC in buried volcanic soils can
also be attributed to high fertility soils typical of many volcanic parent materials (Torn et al., 1997).

Bioturbation by cicadas, earthworms, and small mammals can transport fresh OM to the mineral subsoil (Don et al., 2008; Wilkinson et al., 2009), where the potential for stabilization increases. Bioturbation can also increase the exchange of soil gases and water through the promotion of preferential flow paths, thus potentially reducing the stability of deep SOC by reconnecting buried soils with modern surface conditions. Post-burial additions of OC from plant roots extending into deep soils (Davidson et al., 2011) can facilitate microbial decomposition of ancient SOC (Fontaine et al., 2007) and alter the composition of buried soils (Gocke et al., 2010).

The activity of microbial decomposers in deep and buried soils is poorly understood. Isolation from fresh OC inputs, low nutrient concentrations, and the great spatial patchiness of substrates can lead to reduced microbial biomass and a decrease in functional diversity with increasing soil depth (Ekschmitt et al., 2008; Fierer et al., 2003). Microbiological biomass from soils deep within Bronze Age burial mounds and an underlying paleosol several meters below the surface were 10–20 times lower than those in a nearby arable topsoil (Thomsen et al., 2008). Respiration rates of incubated soils from burial mounds responded to addition of a labile C substrate (glucose) with a two to three week lag compared to modern surface soils (Thomsen et al., 2008).

Microbial transformations of SOC post-burial can also contribute to the persistence of buried SOC. The role of microbial processing in the formation of stable SOC, through preferential sorption of microbial products to mineral surfaces or through occlusion within micropores, is increasingly being recognized (Knicker, 2011; Miltner et al., 2012; Schmidt et al., 2011), especially in deep soils (Liang and Balser, 2008). A study of burial mounds in the central Russian Plain steppe found that the contribution of microbial biomass lipids to the bulk SOC pool increased with buried soil depth and age of the burial mound (Khomutova et al., 2011).

4.4. Effects of parent material on SOC persistence in buried soils

Buried SOC content and persistence are affected by the type of parent material of the buried soil and of the overlying and underlying layers within a profile. A soil’s source material and mineralogy affect its particle size distribution and texture-related mechanisms for SOC stabilization (Fabrizi et al., 2008; Hassink et al., 1997; Jenny, 1941; Krull et al., 2003). Clay content is strongly correlated with SOC storage in deep soil layers (Jobbágy and Jackson, 2000). SOC content and radiocarbon age are strongly influenced by type of mineral composition, which can vary with parent material as well as weathering stage (Torn et al., 1997). Buried tropical volcanic ash soil horizons with poorly crystalline minerals stored more SOC with slower turnover rates than similar horizons containing more crystalline minerals, such as feldspars and gibbsite (Bäsele-Doelsch et al., 2005).

The depth, structure, and composition of material above and below buried soils can affect the diffusivity of oxygen and soil moisture conditions. Soil texture also affects water retention of deep soils (Sotta et al., 2007), influencing soil CO2 efflux and microbial activity. The presence of an impermeable Fe layer under a paleosol buried under a 7-m tall 3330-year-old burial mound in Denmark led to the formation of a water-logged basin within its interior, affecting the decomposition of buried SOC (Thomsen et al., 2008). Soil development of overlying sediment over time, especially in coarser sediments such as loess deposits, can alter soil hydraulic properties (Beerten et al., 2012) and the connectivity of the underlying buried soil with the atmosphere.

4.5. Timescale effects on buried SOC

SOC storage in buried soils is affected by the timescale of the burial process as well as the amount of time since burial. The burial of a surface soil by a depositional material can happen instantaneously, such as during a volcanic eruption or glacial lake dam rupture. Soil burial can occur repeatedly over long periods of time, such as with sequences of volcanic eruptions (Sedov et al., 2001), gradual loess deposition over thousands of years (Feggstad et al., 2004; Zech et al. 2013), and river overbank deposits from repeated floods (Daniels and Knox, 2005).

The rate of sediment accumulation can affect the number of buried horizons in a profile. Very slow deposition rates are more likely to contribute to the formation of thicker surface soil A horizons (Daniels, 2003), while soil burial is more likely to occur when rates of sediment deposition exceed that of soil development. A vertical sequence of multiple buried soils is representative of intermediate frequency of sediment deposition, which allow for pedogenesis in between burial events.

The type of burial process can also affect the timescales of SOC persistence and preservation. A study comparing the reliability of radiocarbon dating of buried SOC concluded that risk of OC contamination was greatest in loess deposits due to more gradual accumulation than in alluvial and flood deposits where burial can occur relatively quickly and where an individual event will tend to deposit thicker sediments (Orlova and Fanychev, 2006). In a study of Japanese Andisols, estimates of past OC inputs based on phytoliths and other SOC measurements suggested that the high levels of C buried by successive tephras deposits had not substantially decreased over the thousands of years since burial (Inoue et al., 2000).
leaching can lead to the reduction of SOC pools in buried soils. Changes in environmental properties regulating the decomposition and persistence of SOC in buried soils can occur through changes in the structure of overlying horizons that alter some of the mechanisms that had previously isolated the buried soil from surface conditions or by exposure to the atmosphere through lateral and vertical erosion. Gully erosion in the Drakensberg escarpment foothills in South Africa has revealed a sequence of paleosols in colluvial sediment (Temme et al., 2012). The existence of multiple sequences of buried horizons in loess deposits extending across a large area of the U.S. Central Great Plains (Miao et al., 2007) became known to researchers through exposure from road construction.

The sensitivity of buried SOC to changing environmental conditions varies with the properties of the mineral matrix and the organic matter, but recent research suggests that deep, ancient SOC can respond to disturbance on annual to decadal time-scales (Devine et al., 2011; Harrison et al., 2011; Veldkamp et al., 2003). In buried soils where OC has been preserved due to isolation from conditions favorable for microbial respiration, exposure to the atmosphere and inoculation with microbial biomass is expected to lead to large losses of SOC. Wetting of paleosols collected below 1 m under 4000-year old burial mounds resulted in comparable respiration rates to those of modern surface soils in the arid Russian steppe (Demkina et al., 2010). Deep SOC pools have been shown to be more responsive to increased temperature during experimental manipulations than surface soil horizons (Schwendemann and Veldkamp, 2006). On the other hand, buried SOC that may have been biochemically altered before or during the burial process (such as during a fire) or since burial may not respond to environmental change as rapidly (Chaopricha, 2013).

The effects of climate change on buried SOC are complex and poorly understood. Changing precipitation patterns coupled with landscape disturbance could alter hydrologic flowpaths and increase soil erosion. These processes can reconnect deep SOC with biologically active surface soil horizons and lead to the exposure of buried SOC to the atmosphere. Buried peatlands, which store considerable amounts of OC, are vulnerable to changes in drainage that accelerate decomposition and increase susceptibility to fire, especially in areas where fire is used to clear vegetation (Limpens et al., 2008). Changes in hydrologic processes or erosion rates that disrupt the accumulation of castic sediments on floodplain peat can also minimize the protection of buried peats to fire (Ellery et al., 2012).

The effect of rising temperatures on buried and deep permafrost C pools has received great attention. Zimov et al. (2006b) estimate that rising temperatures threaten 500 Pg of SOC in Siberian and Alaskan loess permafrost. Estimates of C release from thawing northern permafrost vary widely, partly due to uncertainties in SOC stocks, thawing rates, and the temperature sensitivity of decomposition (Tarnocai et al., 2009). Some authors estimate that permafrost SOC, which has accumulated over tens of 1000s of years, may be released to the atmosphere within a century (Zimov et al., 2006b). Others suggest that increased cryoturbation in thawing permafrost could lead to greater stabilization of SOC, in part by transporting OC to deeper, colder layers with depressed microbial activity (Bockheim, 2007; Kaiser et al., 2007).

Less is known about the response of buried SOC to changes in land management or land use, although recent work suggests that deep SOC stocks are more sensitive than previously thought and can show opposite responses than surface horizons (Baker et al., 2007; Li et al., 2010; Veldkamp et al., 2003). The conversion of conventionally tilled agricultural land to managed successional forests in the U.S. state of Georgia affected total SOC content only in the top 5 cm of soil but resulted in an accumulation of C in particular form in soils up to 2 m depth (Devine et al., 2011). Changes in land use and agricultural practices or hydrology that distribute fresh C down the soil profile can prime microbial activity and stimulate the decomposition and release of deep ancient SOC (Fontaine et al., 2007). Buried soil horizons, where microbes are often C-starved, may show greater responses to priming than modern surface soils (Zhuravleva et al., 2012). More research is needed on how land use and other environmental changes affect deep and buried SOC under different ecosystems (Harper and Tibbett, 2013).

6. Implications of soil C burial for modeling and management

A more explicit incorporation of processes contributing to the delivery, accumulation, and turnover of SOC in the subsoil is fundamental for improving predictions of belowground SOC pools to environmental change (Schmidt et al., 2011). A growing body of evidence suggests that SOC in deep and buried soils can become a large source of ancient C to the atmosphere if reconnected to the atmosphere, and some of it may already be cycling more actively than previously considered (Devine et al., 2011; Fontaine et al., 2007; Gurwick et al., 2008b; Trumbore, 1997). Despite the large stocks of OC in deep soils and their potential sensitivity to climate and land management, models of the effects of climate change in the range, most biogeochemical models include only surface or shallow soil C dynamics. Recent improvements to biogeochemical process models include a vertical dimension to C pools and transformations, the consideration of vertical mixing such as bioturbation, and the input of OC at various depths, rather than the trickle down of OC from surface horizons (Koven et al., 2013).

Given the importance of burial to long-term SOC storage, the incorporation of depositional processes across the landscape into models of SOC should improve predictions of whole-soil profile C dynamics as well as regional C budgets. In Central Europe, rates of SOC burial at the landscape level caused by post-Neolithic agricultural erosion were estimated to be about 170 times greater than regional OC storage in lake and reservoir sediments, but these processes are not included in continental C budgets (Hoffmann et al., 2013).

Our understanding of geomorphic controls on SOC has been greatly improved due to recent research quantifying lateral and vertical C fluxes during erosion across topographic sequences in both cultivated and non-cultivated landscapes and measuring processes leading to the release and protection of C at erosional and depositional sites (Berhe et al., 2012; Doetterl et al., 2012; Van Oost et al., 2012; VandenBygaart et al., 2012) coupled with models of sediment transport (Pelletier et al., 2011; Yoo et al., 2005). To accurately represent buried SOC dynamics, the following additional information should be incorporated into vertically-resolved soil biogeochemistry models: the type of depositional process (e.g., alluvial, colluvial, volcanic, aeolian, anthropogenic, etc.), the rate of burial, the amount of time since burial, the amount of SOC transported in the sediment, the mineral and particle-size composition of the aggrading sediment, and depths of the soil and sediment layers. This information provides a first approximation of the types of mechanisms contributing to the accumulation or loss of SOC at different depths. Inclusion of a process-based understanding of how environmental properties affect SOC decomposition and stabilization in different soil layers in soil C models (Schmidt et al., 2011) will improve predictions of the response of deep and buried SOC to future landscape disturbances driven by changes in climate or human activities. The incorporation of physical mixing of SOC into deeper soil horizons simulating cryoturbation in a land surface model improved estimates of high-latitude SOC stocks down to 3 m (Koven et al., 2009). Even still, lower model predictions than other permafrost pool estimates (Tarnocai et al., 2009).
were attributed by the authors to the exclusion of alluvial and loess depositional processes which lead to the accumulation of substantial deep SOC (Koven et al., 2009). Burial processes are better represented in biogeochemical models of oceans, where environmental conditions play an important role in the fate of organic matter deposited in marine sediments (Kriest and Oschlies, 2013).

From a management perspective, buried SOC pools can be protected through practices that inhibit their exposure to the atmosphere. Alternatives to soil tillage and drainage processes that threaten currently buried SOC pools and that are less disruptive to the soil profile may prevent mobilization and losses of deep and buried SOC. Soil micro-environmental conditions and hydrology could also be managed to stabilize SOC by considering crop choice, rooting depths, and crop residue retention. Assessments of management practices on SOC need to consider the potential presence of large buried SOC pools and the likelihood that the mechanisms contributing to their persistence could be disrupted under different land use and climatic scenarios. The inclusion of geomorphic processes into knowledge of a site’s history can help predict the potential for the presence of buried soils at depth in a landscape.

While exposure of buried soils can result in SOC losses to the atmosphere initially from increased microbial respiration (Demekina et al., 2010); exposed organic horizons can become an important source of nutrients for local primary production. The establishment of vegetation on the exposed soil surfaces can lead to SOC retention, following the successional model of Vitousek and Reiners (1975). The economic importance of paleosols and the role of Quaternary history in shaping soil functional properties that make up the terroir concept are very well recognized (Costantini et al., 2012). More in-depth investigations of soil landscape history are expected to reveal important contributions of buried soils to agricultural productivity and terrestrial C storage.

7. Conclusions

Buried soils have historically been studied to provide insights into past climate change and human environments and to reconstruct landscape evolution. Deeper understanding into the geomorphic, pedogenic, and biological processes that contribute to the accumulation and persistence of SOC in buried soils can also shed light into mechanisms regulating the response of SOC pools to environmental change. Few studies have quantified SOC stocks in paleosols and buried modern soils; however, those that do reveal that substantial SOC stocks, comparable to those of surface soils, may exist in deeply buried soil horizons beyond depths included in most regional or global C inventories. Deep buried soils store SOC of different ages and depths around the world in a variety of different landscapes, climates, and vegetation and altitudinal zones. Organic matter can be buried through a variety of volcanic, aeolian, alluvial, colluvial, glacial, and anthropogenic processes acting independently or in combination. These burial processes differ in their geographic extent, the rate of burial, and their effects on the quantity and turnover of SOC transported in sediment and photo-synthesized at the depositional site.

The incorporation of depositional processes and landscape disturbance events, as well as process-based understanding of how state factors influence the fate of buried SOC, into biogeochemical models coupled to land surface models will improve predictions of the response of large terrestrial OC reservoirs to environmental change and of feedbacks to climate change. There is a rich literature on paleosols, which is primarily descriptive in nature. Increased data on the bulk density of buried horizons and estimates of their geographic extent can improve quantification of the contribution of geomorphic processes to C storage.

Soil burial can isolate OC from the atmosphere, creating environmental conditions unfavorable to microbial decomposition, such as low moisture or low oxygen, which can lead to the persistence of SOC relatively unchanged for thousands of years. Once exposed to ambient conditions, soil microbial activity can recover to rates comparable to surface soils. In other buried soils, biochemical transformations from fire before burial or microbial processing post burial lead to the accumulation of SOC with old radiocarbon ages.

Accelerated landscape disturbance by climate change and human activities can increase the vulnerability of SOC that has been protected from mineralization in deep horizons and increase C losses to the atmosphere. At the same time, increases in erosion and shifts in depositional patterns, resulting from changing climates and extreme weather patterns, could bury higher amounts of OC in the future. Our results highlight the need for landscape-level evaluations on the effects of geomorphic processes on mechanisms of SOC stabilization and incorporation of soil burial processes into soil biogeochemical models.

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