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Organic carbon pools and genesis of alpine soils with permafrost: a review

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Abstract

Soils with mountain permafrost occupy 3.5 million km^2 worldwide, with 70% in central Asia. High-mountain environments have "warm" permafrost, with surface permafrost temperatures of –0.5 to –2 °C and deep active layers (2 to 8 m). From a global database of 41 sites and 312 pedons, alpine soils with permafrost are strongly acid (pH = 5.0 to 5.5), have intermediate cation-exchange capacities (20 to 25 cmol$_c$/kg) and base saturation (44% to 85%), and commonly have an isotic mineral class. Soil organic carbon is concentrated in the upper 30 to 40 cm, with profile density averaging 15.2 ± 1.3 kg m^{-2} (range = <1.0 to 88.3 kg m^{-2}), which is comparable to temperate grasslands (13 kg m^{-2}) but substantially less than moist arctic tundra (32 kg m^{-2}). Mountain soils with permafrost contain 66.3 Pg of soil organic carbon (SOC), which constitutes 4.5% of the global pool. In contrast, the SOC pool in the Arctic is 496 Pg (33% of the global pool). Alpine soils with deep active layers contrast strongly with high-latitude soils in areas of continuous permafrost. Permafrost in the upper 2 m induces cryoturbation in the profile, acts as a barrier to water movement, and generates cooler temperatures resulting in greater SOC levels. High-elevation and high-latitude soils are experiencing warming of air temperature and permafrost and a thickening of the active layer.

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Introduction

Permafrost is defined as a condition whereby a material remains below 0 °C for two or more years in succession (van Everdingen, 1998). Mountain permafrost has received attention from the International Permafrost Association "Mapping and Modeling of Mountain Permafrost" working group. The first map of mountain permafrost and geocryological types was prepared by Gorbunov (1978) from a review of the literature and his experience in the former Soviet Union. The most recent map of mountain permafrost was prepared by Haeberli et al. (1993), who revised Gorbunov's map of mountain permafrost from a review of the literature in response to the concern of climate change impacts on mass wasting. Their map suggested that mountain permafrost occupies 4.88 million km^2, which is nearly double that from the map of Gorbunov (1978).

Gruber (2012) derived a high-resolution estimate of global permafrost distribution by country from a model based on mean annual air temperature (MAAT) and the combined effects of snow cover, exposure to solar radiation, and vegetation. Although the study did not distinguish between high-elevation and high-latitude permafrost, the areas for permafrost in mountainous countries at mid-latitudes were more comparable to the estimates of Gorbunov (1978) than those of Haeberli et al. (1993), suggesting that the global distribution of mountain permafrost may be closer to Gorbunov's estimates than those of Haeberli et al. (1993). However, Gruber's (2012) estimates for the area of mountain permafrost in Mongolia (382,000 km^2) are considerably less than those of Gorbunov (1.4 million km^2) and Haeberli et al. (1993) (1.0 million km^2).

Whereas permafrost covers 23 million km^2 at the high latitudes, mountain permafrost covers an area from 2.5 million km^2 (Gorbunov, 1978) to 4.9 million km^2 (Haeberli et al., 1993) and may account for 14% of the permafrost worldwide. Nearly 70% of the mountain permafrost occurs at the "Third Pole," the high mountains of central Asia (Zhao et al., 2012).

The delineation of permafrost in mountain regions has been determined from drilling or road cut observations (Harris and Brown, 1982; Sharkhuu, 2003; Cheng, 2005; Wu et al., 2010; Mutter and Phillips, 2012); geophysical techniques such as electric-resistivity tomography (ERT), ground-penetrating radar (GPR), and shallow seismic refraction (SSR) (Leopold et al., 2010); as well as inferences from mean annual air temperatures (Péwé, 1983); periglacial features resulting from the presence of an active layer, such as active ice-cemented (lobate) rock glaciers or ice-cored moraines (Urdea, 1998; Fukui et al., 2007; Lilleøren and Etzelmüller, 2011; Alonso and Trombotto, 2012); basal temperature of snow (Lewkowicz and Ednie, 2004; Julián and Chueca, 2007; Ruiz and Trombotto, 2012; Bonnaventure and Lewkowicz, 2013); and models based on climate, topography, and other factors (Harris, 1986; Burn, 1994; Gruber and Hoelzle, 2001; Janke, 2005; Etzelmüller et al., 2007; Allen et al., 2008; Etzelmüller and Frauenfelder, 2009; Arenson and Jakob, 2010; Boeckli et al., 2012; Janke et al., 2012; Zhao et al., 2012). Although permafrost tends to be sporadic in many mid-latitude mountain ranges, it can be discontinuous (i.e., covers 50–90% of ground) or even continuous (90%–100%) at the higher latitudes and in large mountain massifs such as the Qinghai-Tibet Plateau (QTP).

An examination of the topic "alpine soils" in the Web of Science reveals that interest in alpine soils has increased sharply from three papers per year during 1960 to 1990 to 230 publications per year during 2002 to 2012. High-mountain soils are globally important because they are critical sources of water, are centers of biodiversity, and influence global atmospheric and cryospheric systems. Mountain soils with permafrost store large quantities of soil organic carbon (SOC) (Wang et al., 2002; Garcia-Pausas et al., 2007; Ohtsuka et al., 2008; Baumann et al., 2009; Celi et al., 2010; Wu et al., 2010; Budge et al., 2011; Kabala and Zapart, 2012; Dörfer et

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al., 2013; Zollinger et al., 2013) and, therefore, may release large amounts of carbon dioxide in a warming scenario (e.g., Schuur et al., 2013).

The objectives of this study are to (1) update the calculation of the area of mountain permafrost from literature published since 1993 when the last mountain permafrost map was produced, (2) analyze data from mountain soils with permafrost worldwide, (3) determine the SOC contents of these soils, (4) evaluate the role of various soil-forming factors in pedogenesis of soils containing mountain permafrost, and (5) compare and contrast soils influenced by mountain permafrost with those affected by high-latitude permafrost.

In this study, we use the terms *mountain* and *alpine* interchangeably with regard to soils underlain by permafrost, recognizing that *mountain* could be interpreted as any large landform in the form of a peak, with or without permafrost, and *alpine* refers to the area above treeline; however, permafrost may exist in forests at the upper limit of treeline, that is, in the subalpine life zone.

Methods

STUDY AREA

For purposes of this study, we have followed the protocol of Gorbunov (1978) in setting the elevation for mountain permafrost at an arbitrary threshold of 500 m so as to minimize the inclusion of high-latitude permafrost in our estimates. The study area includes the Brooks Range, Rocky Mountains, Coast Ranges, Cascade Range, Sierra Nevada, and Appalachian Mountains of North America; the Andes Mountains of South America; the mountains of Iceland, Greenland, Svalbard, and Fennoscandia; the Alps, Pyrenees, Carpathians, and Urals of Europe; the Caucasus, Himalayan-Karakoram-Hindu Kush (Qinghai-Tibet Plateau), Pamir-Tien Shan-Djungar Alatau, the Khingai-Altai Mountains of central Asia; the Yablonoi-Sayan-Stanovai Mountains of Siberia; the Japanese Alps; and the Southern Alps of New Zealand (Fig. 1). We recognize that the existence of sporadic permafrost in many mountain environments makes area estimates and other interpretations problematic.

The elevation at which mountain permafrost is reported in the literature ranges from as low as 500 m (by the earlier definition) in high-latitude environments such as Iceland, Greenland, Svalbard, and the subpolar portions of the Caucasus Mountains to >5000 m in the central Andes, Qinghai-Tibet Plateau (QTP), and tropical mountains (Table 1). The active-layer thickness ranges from >0.5 m in high-latitude mountain environments such as Iceland or Greenland to more than 8 m in the Andes, European Alps, and Altai Mountains. Permafrost may exist in mountains where the mean annual air temperature (MAAT) is as warm as 1.4 °C, but a value of –3 °C or lower is more typical of areas containing mountain permafrost (Lewkowicz and Ednie, 2004; Etzelmüller et al., 2007; Gruber, 2012). At the other extreme, MAAT values as low as –10 °C have been recorded in mountains of Alaska, Fennoscandia, and the QTP (Table 1). The mean annual precipitation of areas with mountain permafrost ranges from 250 mm yr^{-1} for the Yukon Territory of Canada and parts of the central Asian mountains to over 2000 mm yr^{-1} in the European Alps, Japanese Alps, and the Southern Alps of New Zealand (Table 1). Annual snowfall ranges from a meter to more than 20 m, although wind redistribution can produce significant local differences in snow depth (Table 1).

DISTRIBUTION OF MOUNTAIN PERMAFROST

The distribution of mountain permafrost was determined from regional maps published since the Haeberli et al. (1993) map, including the circumarctic map of Brown et al. (1997) and maps of permafrost in the European Alps (Boeckli et al., 2012), Iceland (Etzelmüller et al., 2007), Norway (Lilleøren and Etzelmüller, 2011), central Asia (Li and Chang, 1996; Marchenko et al., 2007; Zhao et al., 2012), and the Andes (Trombotto, 2000).

PROPERTIES OF ALPINE SOILS

Morphological and analytical data were obtained for soils of each mountain range containing permafrost from the published literature. Our database was derived from 41 studies and included 312 pedons. Data were also obtained from the U.S. Department of Agriculture Natural Resources Conservation Service (NRCS), including Official Soil Descriptions (Soil Survey Division, 2013a), soil classification data (Soil Survey Division, 2013b), and soil characterization data (Soil Survey Division, 2013c).

The profile density of SOC was obtained from the literature or was calculated from data in the literature. The data requirements included classification of soils by either *Soil Taxonomy* (ST) or the World Reference Base for Soil Resources (IUSS Working Group WRB, 2006), SOC or loss-on-ignition (LOI) values, the proportion of coarse fragments (>2 mm), and bulk density. In the database, bulk densities were reported for 19% of the soil horizons, coarse fragments for 43%; 55% of the soils were classified; and 100% of the horizons contained data for SOC or LOI. Where bulk density was not provided, we estimated it from an equation relating SOC or LOI adjusted to SOC to bulk density using the existing database (Fig. 2). It is of interest that this equation has a similar intercept and exponent as those developed for the alpine zone of Mount Mansfield, Vermont, by Munroe (2008).

For pedons where the proportion of coarse fragments was not provided, we used a default value of 45%, which constitutes the mean value from those studies reporting coarse fragments. Where soils had not been classified in ST or WRB, we classified them from existing data, including horizonation, base saturation, and SOC. We used established equations (e.g., Bockheim et al., 2000a) for estimating profile SOC pools.

An SOC budget for high-mountain areas was prepared by summing the product of mean SOC density for each mountain range and the estimated permafrost area of that range. The geomorphology, soils, and climate-change effects in high-elevation environments were compared to those of high-latitude environments from the literature.

Results

WORLD DISTRIBUTION OF MOUNTAIN PERMAFROST

The total area of mountain permafrost was estimated to be 3.56 million km^2 (Table 2), which is intermediate between those of Gorbunov (1978) at 2.46 million km^2 and Haeberli et al. (1993) at 4.88 million km^2 (Table 1). Our data suggest that the single largest region with alpine permafrost is the QTP, at 1.3 million km^2, which constitutes 54% of the plateau. The next largest area of mountain permafrost is Khangai-Altai Mountains of Mongolia and Russia at 1.0 million km^2, Alaska's Brooks Range (263,000 km^2), the Siberian Mountains (255,000 km^2), Greenland (251,000 km^2), the Ural Mountains of Russia (125,000 km^2), the Andes Mountains of South America (100,000 km^2), the Rocky Mountains of the United States

FIGURE 1. Location of mountainous regions of the world considered in this study (see also Table 1).

and Canada (100,000 km^2), and the Fennoscandian mountains (75,000 km^2). The remaining mountain ranges collectively contain less than 100,000 km^2 of alpine permafrost.

PROPERTIES OF ALPINE SOILS UNDERLAIN BY PERMAFROST

The following interpretations are based on the full database of 312 pedons; 14 representative pedons are included in Table 3 from mountain ecosystems throughout the world. Investigation of this data set reveals a number of properties shared by the majority of soil pedons containing mountain permafrost. For instance, although calcareous alpine soils have been reported with pH values as high as 8.0–8.5 (Nimlos and McConnell, 1965; Knapik et al., 1973; Kabala and Zapart, 2012), most alpine soils with permafrost generally are strongly acidic (pH = 5.0 to 5.5). Alpine soils generally have an intermediate cation-exchange capacity (20 to 25 cmol$_c$ kg^{-1}) and base saturation (44% to 85%), abundant SOC especially in the A horizon (4% to 15%, Table 3).

An isotic mineral class is common in mountain soils with permafrost in many areas, including the Rocky Mountains (Knapik et al., 1973; Dahms et al., 2012), Andes (Miller and Birkeland, 1992; Mahaney et al., 2009), Iceland (Arnalds, 2008), Svalbard (Kabala and Zapart, 2012), the European Alps (Egli et al., 2003, 2006; Dahms et al., 2012), and Mount Everest (Bäumler and Zech, 1994). In the NRCS database, 16 of the 26 alpine soil series (62%) have an isotic mineral class (Table 4).

SOIL ORGANIC CARBON DENSITY

Soil organic C tends to be concentrated in the upper 30 to 40 cm of alpine soils even when they are derived from deep unconsolidated sediments (Bockheim et al., 2000a; Yang et al., 2008; Baumann et al., 2009; Budge et al., 2011; Dörfer et al., 2013). Many alpine soils on the QTP are shallow to bedrock; for this reason, SOC density is often reported there to a depth of 30 cm. Soil organic C is reported in the literature at depth intervals ranging from 0 to 30 cm to 0 to 150 cm, and in two studies at variable depths from the soil surface to bedrock (Table 5). The mean value for profile SOC was 15.2 ± 1.3 kg m^{-2} and the range was <1.0 to 88.3 kg m^{-2}. Soil organic C densities were compared among soil orders using the non-parametric Mann-Whitney test. Histosols contained the greatest amount of profile SOC at 52 kg m^{-2}, followed by Spodosols (27.2 ± 4.5), Alfisols (27 ± 11), Gelisols (20.0 ± 6.0), and Mollisols (18.6 ± 11), and with significantly lower quantities of SOC in Inceptisols (13.5 ± 0.86) and Entisols (6.2 ± 1.1) (Fig. 3).

CLASSIFICATION OF ALPINE SOILS UNDERLAIN BY PERMAFROST

Alpine soils described in the literature that are underlain by permafrost occur in 8 of the 12 orders identified in *Soil Taxonomy* (Soil Survey Staff, 1999, 2010), including (from most to least prevalent) Inceptisols, Entisols, Spodosols, Gelisols, Histosols, Alfisols, Mollisols, and Andisols (Table 5). Andisols, not listed in Table 5, occur in soils of Iceland underlain by permafrost (Thorhallsdottir, 1983; Arnalds, 2008) for which we had no SOC data. Cryorthents are com-

TABLE 1

Site conditions for mountain permafrost throughout the world.

Site No.	Mountain Range	Latitude (°)	Elevation (m)	Active-layer depth (m)	Mean annual Air temp. (°C)	Mean annual precipitation (mm/yr)	Mean annual snowfall (m)	References
	North America							
1	Coast Range (U.S.A., Canada)	51–64N	>1230–2000		–1.1 to –6.0	250–1525	10–24	Bonnavenure and Lewkowicz, 2013
	Rocky Mountains							
2	U.S.A.	37–51 60–70N	>3500	2.0 to >5.0	1.1 to –3.3	635–1400	4–12	Péwé, 1983; Bockheim and Burns, 1991; Leopold et al., 2010
3	Canada	51–60N	>2180 to >3000		–1.0 to –1.7	280–1400	2–12	Brown and Péwé, 1973; Harris and Brown, 1982; Bonnaventure and Lewkowicz, 2013
4	Brooks Range	68–69N	>500	0.3–1.5	–5 to –10	180–390	1–2	
5	Cascade Mountains (U.S.A., Canada)	40–51N	>2100	No Permafrost	3.2 to 3.9	2000–2850	10–17	Bockheim, 1972; Sneddon et al., 1972a
6	Appalachian Mountains	42–45N	>1200–1800	>2	1.4 to –3.0	1800	6–8	Bockheim and Burns, 1991; Munroe, 2008
	South America							
7	Andes Mountains	20–57S	>1500 to >5000	3.0–9.0	1.4–6.3	440	1–5	Trombotto, 2000; Trombotto and Borzatta, 2009; Ruiz and Trombotto, 2012; Alonso and Trombotto, 2012
	North Atlantic							
8	Fennoscandia	63–70N	>1400 to >1800	1.4–3.0	–3.9 to –6.1	450–1200	4–12	King, 1986; Harris et al., 2008
9	Iceland	64–66N	>800 to >1000	0.5–0.6	–3.0 to –1.8	600–1500	2–3	Thorhallsdottir, 1984; Etzelmüller et al., 2007
10	Greenland	62–83N	>36 to >165	0.4–0.9	–9.2	200	2–3	Christiansen et al., 2008
11	Svalbard	77–78N	>30	0.9–1.0	–3.8 to –4.4	200–800	1–3	Isaksen et al., 2001; Kabala and Zapart, 2012
	Europe							
12	Alps	44–47N	>2400 to >3000	0.5–8.0	–3.0 to –5.5	1100–2400	4–15	Harris et al., 2008; Gruber and Haeberli, 2009; Budge et al., 2011; Boeckli et al., 2012; Gruber, 2012; Mutter and Phillips, 2012
13	Pyrenees	42N	>2700		0 to –0.7	1600–2000	12–16	Serrano et al., 2001; Garcia-Pausas et al., 2007; Julián and Chueca, 2007
14	Carpathians	44–48N	>2100		–2.6 to –4.0	850–2000	2–9	Urdea, 1998; Skiba, 2007
15	Urals	50–60N	>550	0.3–0.5	–3 to –7	570–1000	2–4	Ivanov and Medvedeva, 2012; Dymov et al., 2013

TABLE 1

Continued

Site No.	Mountain Range	Latitude (°)	Elevation (m)	Active-layer depth (m)	Mean annual Air temp. (°C)	Mean annual precipitation (mm/yr)	Mean annual snowfall (m)	References
	Asia							
16	Caucasus	41–43N	>2800 to >3000		–3 to –11	500–2500	3–10	Grishina et al., 1993
17	Himalayas-Karakoram-Hindu Kush (Qinghai-Tibet Plateau)	30–40N	>4600 to >5000	1.0–7	–1.1 to –7.1	300–500	2–4	Cheng, 2005; Pang et al., 2009; Zhao et al., 2012; Li et al., 2012
18	Altai Mtns., Mongolia	45–53N	>2400 to >3600	1.0–10.0	–1.0 to –4.0	360	1–5	Sharkuu, 2003; Fukui et al., 2007; Sharkhuu et al., 2007
19	Pamir-Tien Shan-Djunder Alatau	38–44N	>2700 to >3000	0.5–4	1.0 to –3.5	700–2000	3–16	Marchenko et al., 2007
20	Yablonoi-Sayan-Stanovoi Mtns., Siberia	50–52N	>1750	>0.5	–3.1	800	3–10	Chevychelov and Volotovskii, 2007
21	Japanese Alps	44N	>1600 to >2000	1.0–5.5	–2.8 to –3.8	2200	3–11	Matsui et al., 1971; Sone et al., 1988; Fukui, 2003; Aoyama, 2005
	South Pacific							
22	Southern Alps, New Zealand	43–45S	>2230 to >2880	No Permafrost	4.0–7.0	1000–1200	4–15	Allen et al., 2008

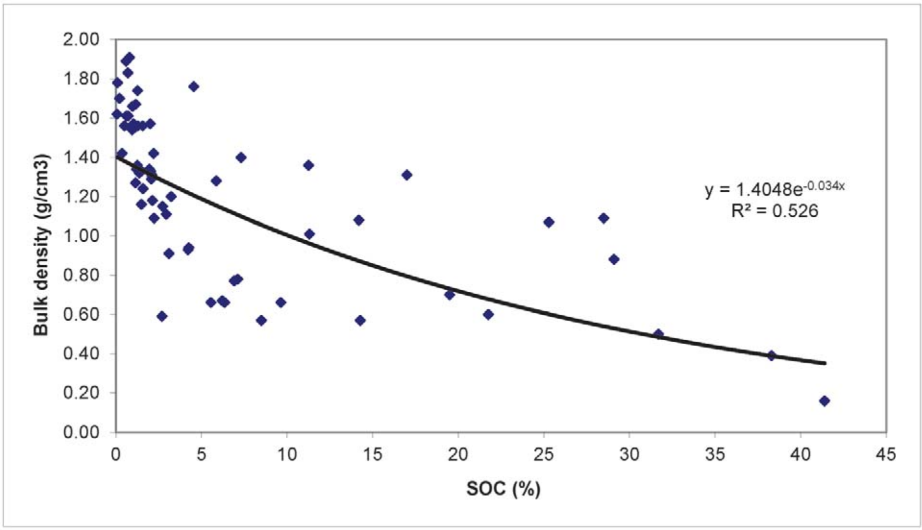

FIGURE 2. Relation of soil organic carbon to bulk density from data contained in the alpine soils with permafrost literature.

mon on young glacial drift, especially in the nival zone (Table 5). Dystrocryepts support alpine grasslands and are common on acidic parent materials. Humicryepts are a dominant soil great group in alpine areas of the world, which are variously identified as Alpine Turf or Mountain Meadows. Haplocryods are restricted to alpine areas with a humid climate, coniferous vegetation, and siliceous parent materials, such as the northern Rocky Mountains and Appalachian Mountains of eastern North America, and the European Alps (Burns, 1990). Gelisols, which require permafrost within 1 to 2 m of the surface, occur primarily in the mountains of Arctic regions, such as in Iceland, Greenland, Svalbard, the Brooks Range in northern Alaska, and the subpolar Caucasus. Histosols are especially common in the alpine zone of the Appalachian and southern Andes Mountains and in local bedrock depressions elsewhere.

The NRCS SSURGO database contains information about more than 23,000 soil series in the United States and its territories. The database contains 11 soil series with likely permafrost and another 15 with possible permafrost (Table 4). These soils are dominantly Humicryepts (46%), followed by Humicryods and Haplocryods (31%), and Dystrocryepts (15%). The soil series with likely permafrost are located mainly in the southern and central Rocky Mountains. The soil series with possible permafrost occur in the Appalachian Mountains, the Rocky Mountains, and the Sierra Nevada.

It is of interest that nearly all of the soils examined in the literature had a cryic soil-temperature regime (STR; mean annual soil temperature 0–7 °C at 50 cm); only a few of the soils had a gelic STR (mean annual soil temperature ≤0 °C at 50 cm). Soils in Gel- suborders and great groups (i.e., have a gelic STR) occur primarily in the Arctic. Gelisols, which must have permafrost within the upper 1 m (Histels and Orthels) or 2 m (Turbels), comprise only 2.35 million km^2 or 52% of the area with alpine permafrost (Bockheim, 2014).

Discussion

FACTORS INFLUENCING SOC DENSITY IN HIGH-MOUNTAIN ENVIRONMENTS

The key factors influencing SOC density of alpine soils underlain by permafrost include (1) vegetation type (Wang et al., 2002; Darmody et al., 2004; Wu et al., 2012), (2) snowfall amount and duration (Burns, 1980; Bockheim et al., 2000a), (3) soil-moisture content (Baumann et al., 2009; Wu et al., 2012; Dörfer et al., 2013), (4) active-layer thickness (Dörfer et al., 2013), (5) elevation (Ohtsuka et al., 2008; Budge et al., 2011), (6) soil type (Matsui et al., 1971; Munroe, 2008; Baumann et al., 2009), and (7) soil age (Birkeland, 1984; Bockheim et al., 2000a; Egli et al., 2001; Darmody et al., 2005; Mahaney et al., 2009; Dahms et al., 2012; Kabala and Zapart, 2012). Factors that were unrelated to SOC density in the literature include aspect (Egli et al., 2006) and the presence of permafrost (at an unspecified depth; Zollinger et al., 2013).

On the QTP, SOC contents varied according to vegetation type, including alpine meadows (54.8–71.6 kg m^{-2}), alpine steppe (11.5–20.8 kg m^{-2}), alpine cold desert (7.8–9.7 kg m^{-2}), and alpine desert (2.2–4.4 kg m^{-2}) (Wang et al., 2002). These differences were attributed to the effects of different vegetation types on litter C production and ecosystem respiration of CO_2.

Snow cover significantly impacts annual carbon and nutrient cycles. Burns (1980) divided ridge-top alpine tundra in the Colorado Front Range into seven topographic-climate regions that explain relative differences in the winter snow cover and SOC contents. Extremely windblown environments and late melting snowbanks had the lowest levels of SOC (4.8–5.5 kg m^{-2}), whereas early melting snowbanks and wet meadows had the highest levels, at 15.1 and 17.0 kg m^{-2}, respectively (Burns, 1980; Bockheim et al., 2000a). These trends likely are related to differences in available soil moisture and its effect on productivity of alpine plant communities and on decomposition rates influenced by growing season length.

Baumann et al. (2009) sampled soils along a 1200-km transect across the Tibetan Plateau, showing that soil moisture was the dominant parameter explaining 64% of the variation in SOC contents. The highest particulate organic matter levels in these soils occur in soils with high soil moisture contents (Dörfer et al., 2013). Another study on the QTP showed that SOC densities were mainly affected by soil depth and soil-moisture content (Wu et al., 2012).

The lower part of the active layer may be an important source of water during summer thawing in permafrost environments. For example, permafrost thawing was identified as a cause of recent hydrologic changes in a Colorado alpine basin (Caine, 2010). Soil organic C stocks

TABLE 2

Distribution of mountain permafrost (10^3 km^2).

Mountain range	Gorbunov (1978)	Haeberli et al. (1993)	Gruber (2012)	This study
North America	~20	400		
Coast Range (U.S.A., Canada)				~37
Rocky Mountains				100
U.S.A.				
Canada				
Brooks Range				263
Cascade Mountains (U.S.A., Canada)				0
Appalachian Mountains				<1
South America				
Andes Mountains	30	270	26	100[a]
North Atlantic				
Fennoscandia		80	71	75[b]
Iceland		50		8[c]
Greenland			250	251[d]
Europe				
Alps	5	50	5	12[e]
Pyrenees				
Carpathians				
Urals				125[f]
Asia				
Caucasus	5	20	2	21[f]
Himalayas-Karakoram-Hindu Kush (Qinghai-Tibet Plateau)	1000	2000	1292	1300[g]
Altai Mtns., Mongolia	1400	1000	382	1000[i]
Pamir, Tien Shan, Djunder Alatau				20[g]
Yablonoi-Sayan-Stanovoi Mtns., Siberia		1000		255[g]
Japanese Alps				<1[h]
North Korea			2.2	
South Pacific				
Southern Alps, New Zealand		10		<1[i]
Total	2460	4880		~3564

[a]Trombotto, 2000; [b]King, 1986; [c]Etzelmüller et al., 2007; [d]Christiansen et al., 2010; [e]Boeckli et al., 2012; [f]Marchenko, unpublished; [g]Li et al., 2012; [h]Fukui, 2003; [i]Allen et al., 2008.

were strongly correlated with active-layer thickness in discontinuous permafrost on the QTP but not in continuous permafrost (Dörfer et al., 2013). The latter authors suggested that permafrost-affected soils in discontinuous permafrost environments were more susceptible to soil moisture changes due to alterations in quantity and seasonal distribution of precipitation, increasing temperature, and, therefore, evaporation.

Ohtsuka et al. (2008) measured SOC pools in alpine to nival zones along an altitudinal gradient (4400–5300 m) on the Tibetan Plateau. The SOC levels increased with elevation from 4400 to 4950 m in the alpine grassland (1.7 kg C m^{-2}), but decreased with further elevation in the nival zone to 1.0 kg C m^{-2}. These differences were attributed to changes in decomposition activity with elevation. In the Swiss Alps, Budge et al. (2011) obtained a significant correlation between the proportion of labile C in 20-cm-deep soil cores and site elevation.

GLOBAL ALPINE SOIL C POOL

Using our estimate of 3.32 million km^2 for the area of mountain permafrost (Table 1) and the SOC densities in Table 4, we estimate the SOC pool for areas with alpine permafrost to be 66.3 Pg, which is 4.5% of the global SOC pool of 1500 Pg (Fig. 4). In

Table 3

Analytical properties of alpine soils of the world that are underlain by permafrost.

Location	Horizon	Depth (cm)	Silt (%)	Clay (%)	pH	SOC (%)	C:N	CEC (cmol+/kg)	BS (%)	Fe_d (%)	Al_d (%)
Uinta, Rocky	Pedon: U92-7; Typic Haplocryolls (Bockheim and Koerner, 1997)										
Mtns., USA	O	0–5									
	A	5–18	10.1	24.2	5.6	2.6	10	14.6	76		
	EB	18–33	2.6	29.5	5.9	0.9	12	7.6	72		
	Bw	33–79	6.1	33.6	5.7	0.3	14	3.6	58		
	C	79–107	4.3	28	5.5	0.3	13	3.6	56		
Rocky Mtns.,	Pedon: P412A; Typic Haplocryepts (Pawluk and Brewer, 1975)										
Alberta, Canada	A	0–8	57	17	4.8	14.7	17	41.8	30	0.92	0.98
	AB	8–12	56	20	4.9	9	20	32.6	20	1.36	1.59
	2Bw	12–18	48	25	5.4	1.4	11	16.8	37	2.79	0.79
	2BC	18–26	43	27	5.3	0.6	7.9	12.3	54	3.22	0.47
	2C	26	35	31	5.6	0.4	7.2	12.2	76	3.75	0.45
Andes Mtns.,	Pedon: COR15; Typic Haplocryepts (Mahaney et al., 2009)										
Venezuela	A	0–15	15.5	6.7		7.8					
	Bw	15–32	17.5	4.6		3.6					
	Cox	32–54	17.3	2.9		0.33					
	Cu	54	19.5	3.1		0.41					
Karkevagge,	Pedon: A1; Oxyaquic Dystrocryepts (Darmody et al., 2000)										
Swedish Lapland	A	0–6			3.6	8.5		9	31	0.03	0.09
	E	6–10			4.1	3.1		2	37	0.032	0.0887
	Bw	10–33			4.3	1.5		2	41	0.0178	0.0677
	Bg	33–69			4.3	1.6		3	41	0.0147	0.0771
	Cg	69–95			4.3	1.5		2	41	0.0164	0.0691
Italian Alps	Pedon: N5; Entic Haplocryods (Egli et al., 2006)										
	Oe	1–10	46.6	24.3	4.1	15.1	19			1.56	0.26
	BE	10–35	29.0	12.5	4.6	6.93	22			3.32	0.81
	Bhs	35–80	19.4	4.6	5.2	3.88	25			3.32	0.99
	BC	80–100	17.0	4	5.3	1.84	20			2.72	0.66
Carpathian Mtns.	Pedon: Lithic Cryorthents (Pelisek, 1973)										
	Oi	0–6									
	A	6–36			4.4	15	18				
	Cr	35–60			4.6	1					
Himalayas,	Pedon: 3; Typic Humicryepts (Smith et al., 1999)										
Tibet	A1	0–13	32.1	7.5	4.6	6.21	13	16.0	55	1.5	0.2
	A2	13–24	21	4.8	4.6	2.7	11	10.0	39	1.2	0.2
	2Bw	24–43	8.9	0.9	4.8	0.49		2.9	66	0.6	0.1
	2BC1	43–74	11.5	1.8	5.1	0.29		2.0	100	0.6	tr
	2BC2	74–116	11.9	1.3	5.2	0.38		2.1	100	0.7	0.1
	2C	116–135	10.3	0.6	5.2	0.12		1.8	100	0.5	tr
Stanovoi Range,	Pedon: 2-89T; Lithic Humicryepts (Chevychelov and Volotovskii, 2001)										
Siberia	Oi	0–2			5.3	54.1			81		
	A	2–10	10.5	5.2	5.2	16.6	35		70	0.63	0.55

TABLE 3

Continued

Location	Horizon	Depth (cm)	Silt (%)	Clay (%)	pH	SOC (%)	C:N	CEC (cmol+/kg)	BS (%)	Fe_d (%)	Al_d (%)
	BhC	10–18	13	8.6	5	25.6	35		62	0.68	0.55
	R	18–25									
Altai Range,		Pedon: 874; Typic Humicryrepts (Krasnoshchekov, 2010)									
Mongolia	A	0–20	22	8	5.4	7.8	16	11.2	64		
	AC	20–35	23	13	5.2	3.1	15	4.1	79		
	M	45–55	22	13	6.5	0.9	22	2.5	87		
Subpolar Ural Mtns.,		Pedon: 25-09; Typic Aquiturbels (Dymov et al., 2013)									
Russia	Oi	0–5			3.9	41.4	56	66			
	Ag	5–14	36	21	5.2	0.8	12	8.8			
	BCg	14–37	37	22	5.1	0.7	12	9.6			
	Cg	37	30	18	5.2	0.1	2	8.1			
Caucasus,		Pedon: 389; Typic Humicryepts (Molchanov, 2008)									
Russia	A	0–10	23.6	22.8	6.7	12.2		50.7	92		
	Bw	15–25	18.7	31.6	7	7.3		43.4	91		
	BC	35–45	16.8	37.6							
Appalachian Mtns.,		Pedon: 94P0304; Lithic Humicryods (USDA, NRCS)									
New York, U.S.A.	Oe	0–33			4.6	19.4		86.1			
	Bh1	33–41	26	3.3	4.6	31.8	22	114	1	0.8	2.4
	Bh2	41–46	27.2	12.2	5.2	11.6	22	53.8	2	0.5	1.2
Pyrenees,		Pedon: Typic Haplocryepts (Parkinson and Gellatly, 1991)									
France	A	0–10	30	5	6.2	7.6		27.3			
	B	10–20	27	9	6.5	1.5		11.3			
	C	20–30	57	5	6.8	0.7		14.3			
Melville Bay, NW		Pedon: 88; Typic Spodorthels (Jakobsen, 1988)									
Greenland	A	0–1			5.1	3.99				1.88	0.80
	E	1–7			5.4	1.39		12.5		2.26	1.00
	Bhs	7–11			5.6	1.6		10.7		2.07	1.60
	Bs	11–19			5.6	0.81		9.1		1.62	1.40
	Bw	19–40			5.7	0.23		3.5		0.88	0.52
	BC	40–65			5.7	0.44		5.8		1.24	0.94
	Cf	65									
Mean values	A		25.8	12.2	5.12	8.19	17	21.0	57	1.0	0.5
by horizon	E		15.8	21.0	5.0	3.1	17.0	7.4		1.9	0.6
	B		21.2	13.3	5.4	6.2	21.5	22.8		1.2	0.9
	BC		24.1	11.2	5.3	0.7	13.3	6.4		1.7	0.5
	C		24.4	12.7	5.5	0.6	11.1	6.4	72	1.4	0.3

contrast, the circumarctic region contains 496 Pg of SOC to a depth of 100 cm (Tarnocai et al., 2009), which is about 33% of the global SOC pool. However, as much as 60% of the SOC in high-latitude permafrost-affected soils occurs below 100 cm in the near-surface permafrost (Tarnocai et al., 2009).

SOIL-FORMING PROCESSES IN ALPINE SOILS WITH PERMAFROST

In the scheme of Bockheim and Gennadiyev (2000), the dominant soil-forming processes in high-mountain environments

TABLE 4

Soil series in alpine areas of the U.S.A.

Series Name	Series Name	Series Name	Area (km^2)	Other	Order	Suborder	Great Group	Subgroup	Particle-size Class	Mineralogy Class	CEC Class	STR	Other	MAP (mm)	MAAT (°C)	Vegetation
					Likely Permafrost											
MEREDITH	CO	91170	369	UT	Inceptisols	Cryepts	Humicryepts	Spodic Humicryepts	loamy-skeletal	isotic		cryic		890	-3.3	alpine tundra
BROSS	CO	64540	261	WY	Inceptisols	Cryepts	Humicryepts	Typic Humicryepts	loamy-skeletal	isotic		cryic		890	-2.2	alpine tundra
MIRROR	CO	75980	308	WY	Inceptisols	Cryepts	Humicryepts	Typic Humicryepts	loamy-skeletal	isotic		cryic		890	-1.1	alpine tundra
PENITENTE	NM	76800	311		Inceptisols	Cryepts	Humicryepts	Typic Humicryepts	loamy-skeletal	isotic		cryic		1200	-1.1	alpine tundra
TEEWINOT	WY	53280	216	CO, ID	Inceptisols	Cryepts	Humicryepts	Lithic Humicryepts	loamy-skeletal	mixed	superactive	cryic		1000	-1.1	alpine tundra
HENSON	CO	225650	914		Inceptisols	Cryepts	Dystrocryepts	Typic Dystrocryepts	loamy-skeletal	isotic		cryic		990	-0.6	alpine tundra
MORAN	WY	427640	1732	CO, ID	Inceptisols	Cryepts	Humicryepts	Typic Humicryepts	loamy-skeletal	mixed	superactive	cryic		1140	0	alpine tundra
NAMBE	NM	864830	3503		Spodosols	Cryods	Haplocryods	Entic Haplocryods	loamy-skeletal	isotic		cryic		1120	0	spr-fir
LEIGHCAN	WY	775700	3142	CO, UT	Inceptisols	Cryepts	Dystrocryepts	Typic Dystrocryepts	loamy-skeletal	mixed	superactive	cryic		1140	0	spr-fir
WALCOTT	WY	119920	486		Inceptisols	Cryepts	Humicryepts	Typic Humicryepts	loamy-skeletal	mixed	superactive	cryic		1140	0.6	alpine tundra
MAHOOSUC	ME	39530	160		Histosols	Folists	Cryofolists	Typic Cryofolists				cryic	dysic	1250	1.1	bal fir krumm
SURPLUS	ME	297150	1203	NH, NY	Spodosols	Cryods	Haplocryods	Aquic Haplocryods	coarse-loamy	isotic		cryic		1270	1.1	balsam fir
SADDLEBACK	ME	641340	2597	NH, NY	Spodosols	Cryods	Humicryods	Lithic Humicryods	loamy	isotic		cryic		1270	1.1	balsam fir
BEMIS	ME	91100	369		Inceptisols	Aquepts	Cryaquepts	Aeric Cryaquepts	loamy	mixed		cryic	shallow	1270	1.1	spr-fir
ENDLICH	CO	33400	135		Inceptisols	Cryepts	Dystrocryepts	Typic Dystrocryepts	loamy-skeletal	isotic		cryic		1200	1.1	spr-fir
					Possible Permafrost											
COUCHSACHRAGA	NY	100110	405		Spodosols	Cryods	Humicryods	Lithic Humicryods	sandy	isotic		cryic		1400	1.7	spr-fir
ESTHER	NY	28170	114		Spodosols	Cryods	Humicryods	Aquandic Humicryods	medial	amorphic		cryic		1400	1.7	balsam fir
SANTANONI	NY	83110	337		Spodosols	Cryods	Humicryods	Typic Humicryods	sandy-skeletal	isotic		cryic		1400	1.7	spr-fir

TABLE 4

Continued

Series Name	Series Name	Series Name	Area (km^2)	Other	Order	Suborder	Great Group	Subgroup	Particle-size Class	Mineralogy Class	CEC Class	STR	Other	MAP (mm)	MAAT (°C)	Vegetation
SKYLIGHT	NY	215840	874		Spodosols	Cryods	Humicryods	Lithic Humicryods	sandy	isotic		cryic		1400	1.7	spr-fir
Fishsnooze	CA	33720	137		Inceptisols	Cryepts	Humicryepts	Xeric Humicryepts	loamy-skeletal	isotic		cryic		1140	1.7	mtn hem-pine
Sumeadow	CA	73200	296		Inceptisols	Cryepts	Humicryepts	Xeric Humicryepts	loamy-skeletal	isotic		cryic		760	1.7	pines-sedges
ARCHROCK	CO	78510	318		Inceptisols	Cryepts	Humicryepts	Typic Humicryepts	loamy-skeletal	micaceous		cryic		860	2.2	alpine tundra
MUMMY	CO	154400	625		Inceptisols	Cryepts	Humicryepts	Typic Humicryepts	loamy-skeletal	micaceous		cryic		890	2.2	alpine tundra
TRAILRIDGE	CO	167030	676		Inceptisols	Cryepts	Humicryepts	Typic Humicryepts	loamy-skeletal	micaceous		cryic		890	2.2	spr-fir
FALLRIVER	CO	418700	1696		Inceptisols	Cryepts	Dystrocryepts	Typic Dystrocryepts	loamy-skeletal	isotic		cryic		810	3.3	spr-fir
GRENADIER	CO	98890	401		Spodosols	Cryods	Haplocryods	Entic Haplocryods	loamy-skeletal	isotic		cryic		640	3.3	spr-fir
			21585													

CEC = cation-exchange capacity class, STR = soil-temperature regime, MAP = mean annual precipitation, MAAT = mean annual air temperature.

CO = Colorado, NM = New Mexico, WY = Wyoming, ME = Maine, NY = New York, CA = California, UT = Utah, ID = Idaho, NH – New Hampshire.

TABLE 5

Soil organic carbon density in alpine soils underlain by permafrost.

Area	No. of pedons	Elevation (m)	Active layer depth (m)	Vegetation	Soil great group	Depth inter val (cm)	SOC (kg/m^2)	References
Brooks Range, AK	3	>500	>0.6	Dry tundra	Mollorthels, Umbrorthels	0–100	7.5–14.2	Bockheim et al., 1998
Rocky Mts., AB	4	2240–2969	nd	Alpine turf	Cryorthents, Haplocryepts, Humicryepts	0–50	6.6–14.8	Knapik et al., 1973
Rocky Mtns., AB	3	2550	>1.0	Alpine tundra	Haplocryepts	to bedrock (26–45 cm)	6.9–8.4	Pawluk and Brewer, 1975
Rocky Mtns., MT	3	>3000	>1.0	Alpine tundra	Humicryepts, Cryorthents, Cryaquepts	0–100	7.2–38.1	Nimlos and McConnell, 1965
Rocky Mtns., CO	6	>3500	1.5–2.0	Alpine tundra	Haplocryepts, Dystrocryepts?	0–100	9.1–24.4	Retzer, 1974
Rocky Mtns., CO	8	>3500	>1.0	Alpine tundra	Cryorthents, Dystrocryepts, Haplocryalfs	0–100	0.58–21.3	Birkeland et al., 1987
Rocky Mtns., UT	12	3518–3735	>2	Alpine tundra	Dystrocryepts, Haplocryolls, Haplocryalfs, Argicryolls	0–100		Munroe, 2007
Rocky Mtns., UT	12	3385–3700	>2	Alpine tundra	Haplocryalfs, Dystrocryepts, Cryorthents, Humicryepts, Haplocryolls	0–100	3.0–22.6	Bockheim and Koerner, 1997
Appalachian Mtns., VT	33	1337–1917	nd	Alpine tundra	Cryofolists, Cryorthents, Dystrocryepts	to bedrock (23–50 cm)	11.3–37.0	Munroe, 2008
Appalachian Mtns., NH	8	>1500	>1.0	Alpine tundra	Haplocryods, Dystrocryepts, Cryaquepts, Cryosaprists	0–100	18.6–67.6	Bliss, 1963
Appalachian Mtns., ME	2	1335	>1.0	Alpine tundra	Haplocryods	0–100	28.4–31.6	Bliss and Woodwell, 1965
Venezuelan Andes	3		>1.0	Alpine tundra	Dystrocryepts, Haplocryepts?	0–100	12.6–16.4	Mahaney et al., 2009
Peruvian Andes	6	~4000	>1.0	Alpine tundra	Haplocryalfs, Haplocryods, Cryorthents, Haplocryepts	0–>70	5.2–88.3	Miller and Birkeland, 1992
Storbreen, Norway	19	1300–1450	>1.0	Alpine foreland	Cryorthents	0–50	1.0–14.0	Darmody et al., 2005
Swiss Alps	3	2000–2050	>1.0	Alpine tundra	Haplocryods	0–100	17.3–24.1	Egli et al., 2001
Swiss Alps	5	2285–2653	nd	Alpine grassland	nd	0–30	5.5–10.2	Budge et al., 2011
Swiss Alps	11	2569–2695	nd	Alpine grassland	Haplocryods?	0–30	6.1–13.2	Zollinger et al., 2013
Italian Alps	10	1200–2420	nd	nd	Humicryepts, Haplocryepts, Haplocryods	0–100	7–35	Egli et al., 2006
Pyrenees	35	1845–2900	nd	Alpine grassland	nd	to bedrock (13–74 cm)	5.9–30	Garcia-Pausas et al., 2007
Pamirs	2	2580	>0.7	Alpine tundra	Humicryepts?	0–100	11.3–14.6	Karavayeva, 1958

TABLE 5

Continued

Area	No. of pedons	Elevation (m)	Active layer depth (m)	Vegetation	Soil great group	Depth inter val (cm)	SOC (kg/m^2)	References
Pamirs	5	3600–4150	>1.0	Alpine tundra	Haplocryepts?	0–80	3.5–6.0	Kann, 1965
Tien Shan	8	3300–3600	>1.0	Alpine meadows	Humicryepts?	0–80	7.4–20.6	Rubilin and Dzhumagulov, 1977
Qinghai-Tibet Plateau	11	4300	4-7	Alpine meadow	Haplorthels?	0–30	3.4–10	Dörfer et al., 2013
Qinghai-Tibet Plateau	38	nd	<0.8–>4	Alpine steppe	Haplocryepts, Gelorthents?	0–30	9.3–11	Wang et al., 2008
Qinghai-Tibet Plateau	405	nd	nd	Alpine meadow, steppe	nd	0–100	4.4–9.5	Yang et al.,2008
Qinghai-Tibet Plateau	24	4400–5300	nd	Alpine meadow	nd	0–30	2.6–14	Ohtsuka et al., 2008
Qinghai-Tibet Plateau	62	nd	nd	Alpine meadow	Haplocryepts?	0–75	29–53	Wang et al., 2002
Qinghai-Tibet Plateau	62	nd	nd	Alpine steppe	nd	0–30	9.0–16	Wang et al.,2002
Khangai-Altai, Mongolia	5	>2100–>2600	0.19–0.55	Alpine tundra	Gelorthents, Humigelepts, Aquorthels	to bedrock (25–50 cm)	13.5–38.0	Krasnoshchekov, 2010
Ural Mtns.	10	>550	0.3–0.5	Mtn. tundra	Aquorthels?	0–50	7.7–39.3	Dymov et al., 2013
SE Yakutia	2	1695–1750	nd	Alpine tundra	Haplocryods	to bedrock (18–40 cm)	7.7–15.4	Chevychelov and Volotovskii, 2007

SOC = soil organic carbon.
AK = Alaska, AB = Alberta, MT = Montana, CO = Colorado, UT = Utah, VT = Vermont, NH = New Hampshire, ME = Maine.

with permafrost are andisolization, melanization, podzolization, cryoturbation, paludization, and gleization. Cambisolization was not identified in the paper by Bockheim and Gennadiyev (2000), but was suggested by D. Yaalon (personal communication) as an additional key soil-forming process.

Andisolization refers to the formation of amorphous minerals (allophone) from weathering of volcanic ash and other silica-rich materials, a process that is especially prevalent in high-rainfall environments (Parfitt et al., 1983). High-mountain soils often have an isotic mineral class (Table 3), which is defined on the basis of two criteria: a pH ≥8.4 in a NaF extract and a ratio of 1500 kPa water content to percent clay of ≥0.6 (Soil Survey Staff, 1999, 2010). The test with NaF is designed as a relatively quick measurement of the content of the non-crystalline minerals (e.g., allophone, imogolite) in a soil (Fieldes and Perrott, 1966). The initial pH of the NaF is 7.5–7.8 so that a value in excess of 8.4 is indicative of poorly crystalline minerals as the fluoride anion displaces hydroxyl ions and complexes Al.

There are 1001 soil series with an isotic mineral class in the NRCS database (Soil Survey Division, 2013a, 2013b, 2013c). Fifty-three percent of these soil series are Inceptisols, which also constitute the dominant order of mountain soils with permafrost identified in the present study. Sixty-four percent of the soils containing an isotic mineral class in the NRCS database have a frigid or cryic soil-temperature regime, suggesting that a cryic soil-temperature regime is conducive to the development of amorphous minerals. None of the 52 Gelisols in the NRCS database contains an isotic mineral class; rather, they all have a mixed mineralogy. Together these findings suggest that the formation of amorphous minerals common to high-mountain soils is favored by abundant soil moisture, a frigid or cryic but not gelic soil-temperature regime, and siliceous parent materials.

Melanization refers to the accumulation of well-humified organic matter within the upper mineral soil. This process is evidenced in high-mountain soils by thick A horizons (Table 3) and comparatively high SOC densities (Table 5).

Although not requiring permafrost, cryoturbation (frost-stirring) is a common process in permafrost-affected soils and is manifested by patterned ground on the land surface and irregular and broken horizons, organic matter accumulation on the permafrost table, oriented stones, and silt caps within the soil. Patterned ground is a common feature in high-mountain environments, covering from 5% to 19% of the ground surface (Johnson and Billings, 1962; Niessen et al., 1992; Hort and Luoto, 2009; Feuillet, 2011). In addition to high-latitude mountain environments, such as Iceland (Arnalds, 2008), Svalbard (Kabala and Zapart, 2012), and the Scandinavian Mountains (Darmody et al., 2000, 2004), cryoturbation has been reported in the central Rocky Mountains of the

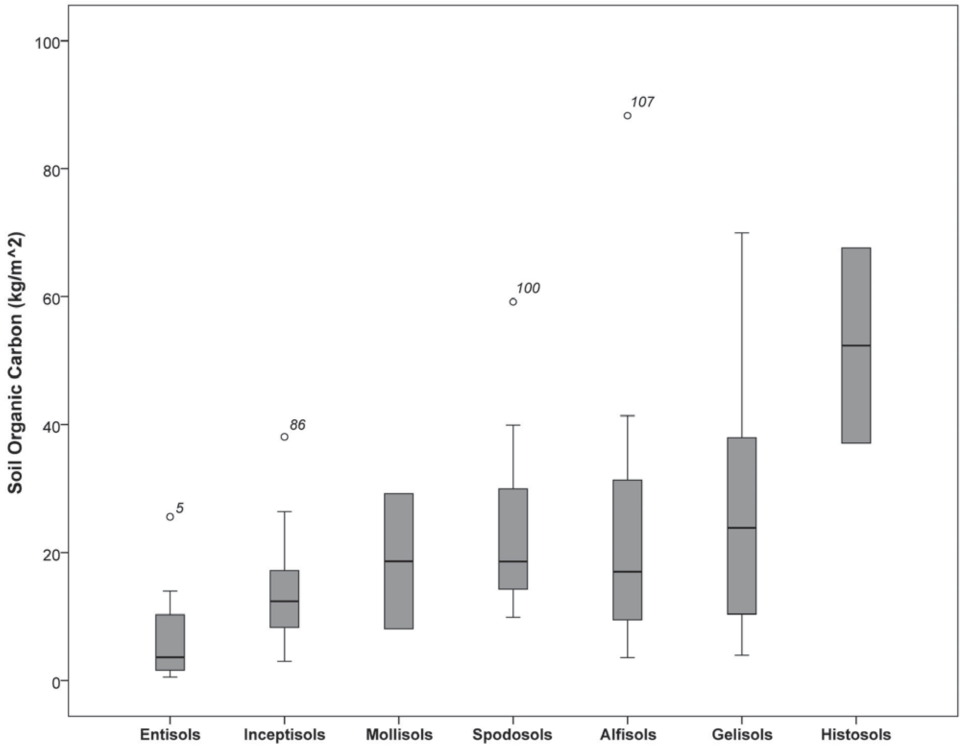

FIGURE 3. Soil organic carbon density in relation to soil taxa.

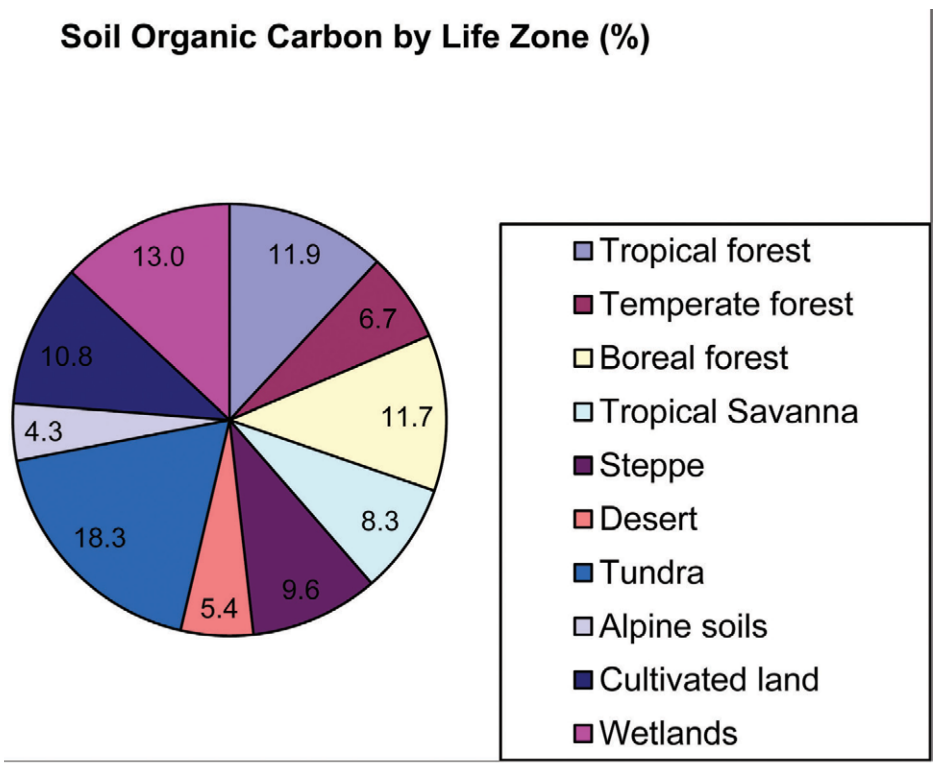

FIGURE 4. Distribution of soil organic carbon in alpine soils (this study) relative to soils of other life zones (Jobbágy and Jackson, 2000).

United States (Bockheim and Koerner, 1997; Munroe, 2007), the Swiss Alps (Celi et al., 2010; Zollinger et al., 2013), and the Ural Mountains (Dymov et al., 2013). In many of these locations, it is not clear whether the cryoturbation is active today or is a relict from colder (glacial?) climates. Although the permafrost table is deep in many alpine soils with permafrost today, the active layer may have been shallower under colder climatic conditions in the past, making these soils function more like Turbels. Outside of the areas noted above, we found only a few published reports of cryoturbation in high-mountain soils of central Asia or the Andes. Smith et al. (1999) reported hummocky, cryoturbated ground in moist depressions around 5000 m on the Qinghai Plateau, but no permafrost was observed within 2 m of the surface. In the Khangai Mountains of Mongolia, Krasnoshchekov (2010) attributed the lack of cryoturbation to the presence of dry permafrost.

Cambisolization leads to the formation of weakly developed Bw (cambic) horizons. This process is pervasive in mountain environments with permafrost, as evidenced by the abundance of Inceptisols (Table 5). The abundance of Inceptisols in alpine regions may be due to the fact that many of these soils are young and pedogenesis is inhibited by slow rates of weathering and horizon formation. Küfmann (2008) suggested that many of the Inceptisols (Cambisols) in alpine karst of the Northern Calcareous Alps of Germany are of eolian origin rather than from weathering in situ. However, this does not imply that Inceptisols in alpine areas worldwide are of similar origin.

Podzolization is a complex collection of processes that includes eluviation of base cations, weathering transformation of Fe and Al compounds, mobilization of Fe and Al in surface horizons, and transport of these compounds to the spodic horizon as Fe and Al complexes with fulvic acids and other complex polyaromatic compounds. This process occurs in many high-mountain environments throughout the world, particularly in subalpine areas with a humid climate, coniferous vegetation, and siliceous parent materials (Burns, 1990; Skiba, 2007).

Gleization refers to redoximorphic features such as mottling and gleying that result from aquic conditions; this occurs in most mountain ranges, especially in bedrock depressions.

Paludization refers to the accumulation of histic materials, but does not necessarily imply poor drainage conditions; it is a dominant process in the alpine zone of the Appalachian Mountains (Munroe, 2008). Many of these soils are classified as Cryofolists, which are saturated for less than 30 consecutive days per year and may store up to 60 kg m^{-2} of SOC.

The other soil-forming processes identified by Bockheim and Gennadiyev (2000) occur to a limited extent in high-mountain environments, including alluviation, biological enrichment of bases, and calcification. Argillic horizons have been reported in the central Rocky Mountains of the United States (Burns, 1980; Bockheim and Koerner, 1997; Munroe, 2007). These features are common in pedons between 3400 and 3600 m in the Uinta Mountains of Utah; however, it is unclear whether they are forming under modern conditions or are relict from a previous milder soil climate (Bockheim and Koerner, 1997; Munroe, 2007).

Haplocryolls with high base levels are common in the central Rocky Mountains (Burns, 1980; Bockheim and Koerner, 1997; Munroe, 2007). Bockheim et al. (2000b) reported that base cycling is enhanced in alpine communities dominated by *Acomastylis* (*Geum*) *rossii*, as evidenced by large profile quantities and high tissue concentrations of Ca. Calcification has been reported in alpine soils in the northern Rocky Mountains (Nimlos and McConnell, 1965; Knapik et al., 1973), Svalbard (Kabala and Zapart, 2012), and in the semiarid mountains of central Asia (Kann, 1965).

COMPARISON OF ALPINE AND ARCTIC SOILS

A question that arises from this review: What role does permafrost occurring at depths of 2 to 8 m play in pedogenesis? To address this question, we compared soils in high-elevation (alpine) environment with those in high-latitude (arctic) environments, recognizing that active-layer thicknesses have changed during glacial-interglacial intervals. The area of high-elevation permafrost is considerably less than that of high-latitude permafrost (Table 6). The mean SOC density is also greater for arctic soils, and as a result the total SOC pools are larger for arctic regions. At the pedon scale, the depth-distribution of SOC in profiles from the arctic and alpine regions varies significantly. In alpine regions with permafrost, SOC is concentrated in the upper 30 to 40 cm of the soil profile and decreases regularly with depth (Bockheim et al., 2000a; Baumann et al., 2009; Yang et al., 2008; Dörfer et al., 2013). In contrast, SOC in arctic soils is concentrated at the base of the active layer and in the transition zone between the active layer and near-surface permafrost, because of intensive cryoturbation accompanied by compaction (Bockheim et al., 2003).

The thermal regime also varies in important ways between arctic and alpine permafrost soils. Most notably, alpine permafrost soils typically feature a deep active layer, which can be attributed to high summer temperatures and reduced volumes of ground ice (Pang et al., 2009). Figure 5 shows measured soil temperatures through a representative year at two depths in an alpine soil at 3,700 m in the Uinta Mountains of the central Rocky Mountains and a Gelisol in arctic Alaska. On the basis of a MAAT <0 °C, permafrost is likely present beneath this alpine soil profile, although the active layer is more than 100 cm thick. Similar conditions occur in the ultra-continental region of central Sakha, Russia, where the active-layer depth is in excess of 2 m despite a MAAT of –9 °C or lower (Lessovaia et al., 2013).

Both soil profiles in Figure 5 exhibit rapid warming in May; however, temperatures in the arctic soil fail to reach 5 °C during the summer, remaining close to 0 °C at a depth of 64 cm. In contrast, summer temperatures at nearly 1 m in the alpine soil profile are almost 5 °C warmer. Warmer summer temperatures are conducive to more rapid breakdown of organic matter, perhaps contributing to the reduced SOC density in alpine permafrost soils.

Temperatures in both profiles fall to ~0 °C by late September, but the arctic soil remains at this temperature longer (until early December) due to a zero-curtain effect involving the freezing of abundant soil moisture that cannot drain from the profile because of ice at depth. This moisture contributes to the formation of ice lenses and other massive ground ice features that promote cryoturbation. As noted earlier, patterned ground produced by cryoturbation has been reported for alpine soils with permafrost, but it is unclear to what extent cryoturbation is occurring in these soil profiles today. Although the active layer typically extends beneath the base of the pedon in these soils, cryoturbation may have been a more active process during colder climates of the Pleistocene when a shallower permafrost table, and (possibly) more significant accumulations of massive ground ice, impeded subsurface drainage, enhancing volumetric changes associated with freeze/thaw cycles. Under modern conditions with a deep active layer and relatively warm summer temperatures in the solum, the presence of permafrost at depths >2 m may still aid cryoturbation closer to the surface by promoting two-way freezing of the active layer. However, beyond this indirect effect, deep permafrost may contribute little to pedogenesis in alpine soils. It is of interest that alpine Gelisols appear to occur only where the MAAT is –5 °C or colder (Table 1).

CLIMATE WARMING IN ALPINE AND ARCTIC SOILS WITH PERMAFROST

Both the high-mountain and high-latitude regions are experiencing major changes due to climate warming. The MAAT in

TABLE 6

A comparison of geomorphology and soils of high-elevation and high-latitude environments.

Parameter	Alpine	References	Polar	References
Area (10^6 km^2)	3.3	This study	22.8	Tarnocai et al., 2009
SOC density (kg/m^2 to 1 m)	6.7–45	This study	32–70	Tarnocai et al., 2009
SOC storage (Pg to 1 m)	55	This study	496	Tarnocai et al., 2009
Patterned ground and/or cryoturbation (%)	5–19	Johnson and Billings, 1962; Feuillet, 2011	10–57	Bockheim et al., 1998; Tarnocai et al., 2007
Active-layer depth (m)	2.0–8.0	This study	0.3–2.0	Mazhitova et al., 2004; Tarnocai et al., 2004
Air temperature warming (°C/decade)	0.5–0.55 (Altai)	Fukui et al., 2007	0.6 (arcticwide)	CRUTEM 3v
	0.8 (QTP)	Li et al., 2012	0.7 (W. Ant. Peninsula)	Turner et al., 2009
	1.0 (Front Range, CO)	Leopold et al., 2010		
	0.06–0.3 (Tien Shan)	Marchenko et al., 2007		
	0.6 (Altai)	Sharkhuu, 2003		
TTOP warming (°C/decade)	0.1–0.2 (Tien Shan)	Marchenko et al., 2007	0.3–0.7 (N. Alaska)	Osterkamp, 2007
	0.1–0.4 (QTP)	Li et al., 2012	1.0 (NVL, Ant.)	Guglielmin and Cannone, 2011
	0.4–0.7 (Svalbard)	Isaksen et al.,2007		
Active-layer depth (cm/yr)	1.3 (QTP)	Li et al., 2012	0 (NE Greenland)	Christiansen et al., 2004
			0 (Eur. Russia)	Mazhitova et al., 2004
			0 (Arctic Canada)	Tarnocai et al., 2004
			1.4 (Sweden)	Christiansen et al., 2010

Notes: SOC = soil organic carbon, QTP = Qinghai-Tibet Plateau, CO = Colorado (U.S.A.), TTOP = temperature top of the permafrost, NVL = North Victoria Land.

high-mountain regions with permafrost has increased from 0.3 to 1.0 °C decade^{-1} (Table 6). In the Arctic, the MAAT has increased 0.6 °C decade^{-1}, and along the western Antarctic Peninsula, it has increased 0.7 °C decade^{-1}. The temperature at the top of the permafrost (TTOP) in mountain environments has increased from 0.1 to 0.7 °C decade^{-1} over the past several decades (Table 6). In the Arctic, TTOP has increased 0.3–0.7 °C decade^{-1}, and in North Victoria Land, Antarctica, it has increased 1.0 °C decade^{-1}. The active-layer depth has increased 1.3 cm yr^{-1} on the QTP and in sporadic permafrost of subarctic Sweden (Table 6). However, the active-layer depth has remained unchanged over the past two decades in areas of continuous permafrost in the circumarctic (Mazhitova et al., 2004; Tarnocai et al., 2004). In soils with excess ground ice, such as many Arctic Gelisols, temperature increases and deepening thaw depths lead to changes in the hydrologic cycle, thermokarst, and a variety of positive feedbacks. It is less clear what the effect of warming soil temperatures will be on alpine soils where the permafrost table lays beneath the base of the soil profile.

Conclusions

- From recent literature, we estimate the area of mountain permafrost to be 3.6 million km^2 which is intermediate between earlier estimates by Gorbunov (1978) and Haeberli et al. (1993).
- The SOC density of areas with mountain permafrost is estimated to be 66 Pg, which constitutes 4.5% of the world total. In contrast, the SOC density of Arctic permafrost is 496 Pg, which accounts for 33% of the global total.
- SOC densities in alpine grasslands are comparable to those of temperate grasslands but are less than those in arctic tundra.
- Estimates of SOC density require the measurement of horizon thickness, bulk density, the concentration of coarse fragments (>2 mm), and SOC. Few of these measurements have been made in mountain permafrost environments outside the Qinghai-Tibet Plateau.
- Estimates of SOC density in mountain environments with permafrost are complicated by the variable depth to bedrock and the abundance of coarse fragments in the soils. SOC tends to be concentrated in the 0–30 cm layer in areas with deep unconsolidated materials.
- The presence of permafrost at depths in excess of 2 m does not appear to directly impact SOC accumulation and other soil properties, although cold temperatures at depth may promote cryoturbation in the solum.

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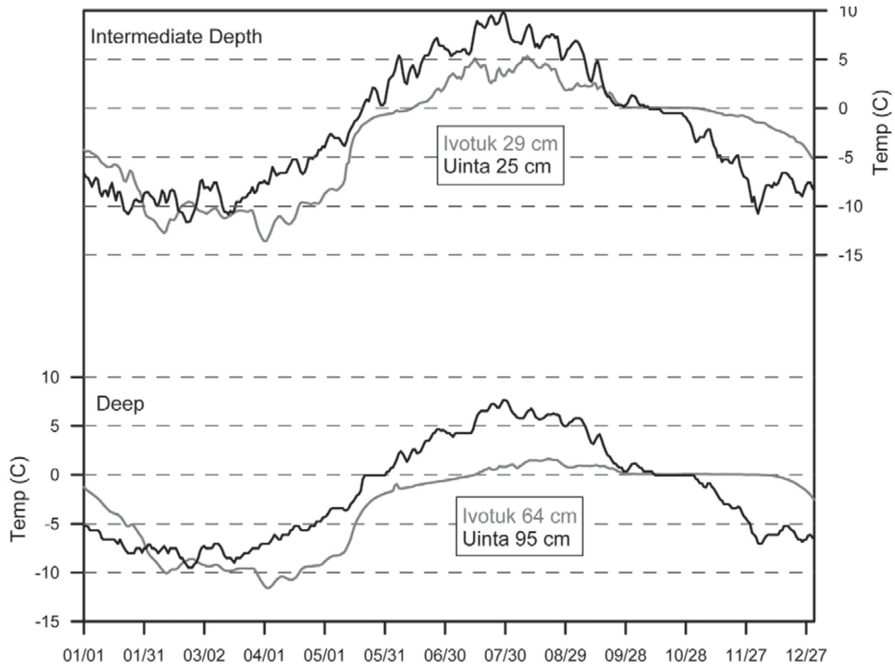

FIGURE 5. A comparison of soil temperatures for alpine tundra in the Uinta Mountains (Munroe, unpublished) with those from arctic tundra (IvotUk, Alaska).

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