



Organic and inorganic carbon storage in soils along an arid to dry sub-humid climosequence in northwest of Iran



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ABSTRACT

The importance of soil carbon in different biomes of the earth is well understood. Most of the studies on terrestrial carbon cycle have been focused on the surface horizon of soils, but deeper soils have rarely been considered. The effects of arid, semi-arid and dry sub-humid climates on soil organic carbon (SOC), soil inorganic carbon (SIC) and soil total carbon storage (STC) along a soil climosequence, on basaltic underlying rocks were investigated. SOC and SIC content and storage showed reverse trends with increasing soil depth. STCs increased by increasing mean annual precipitation (MAP) from 3.75 and 6.28 kg m⁻² in the arid and semi-arid regions, respectively to 11.32 kg m⁻² in the dry sub-humid. Despite lower SOC in the soils of arid region, the highest SICs/SOCs ratio was obtained in the arid climate, which indicates the importance of climate on SIC storage compared to the SOC content. The average times to store SIC in the dry sub-humid, semi-arid and arid regions were calculated as 15,400, 23,100 and 26,000 years, respectively; this indicates that SIC is stored more rapidly in wetter climates due to more weathering. SICs constituted the dominant proportion of STCs which decreased by increasing MAP from in the arid region in comparison with other wetter regions (65%, 74.4% and 84.8% in the semi-arid, dry sub-humid and arid regions, respectively).

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1. Introduction

Carbon (C) storage has attracted a significant amount of attention from researchers (Shi et al., 2012; Zhang et al., 2015), policymakers and environmental scientists in recent decades due to the effect of carbon levels on various aspects of human life and the environment (e.g. reducing greenhouse gases emissions) (Shi et al., 2012; Lal, 2013b). Rising temperature and elevated atmospheric carbon dioxide (CO₂) simultaneously affect the dynamics of soil total carbon (STC) (Wang et al., 2016). There are five main C pools on the earth: (1) the lithosphere, including fossil fuels and sedimentary rock deposits such as limestone, dolomite and chalk (66–100 million Pg); (2) oceans (38,000–40,000 Pg); (3) soil organic carbon (SOC) (1500–1600 Pg) (Lal, 2004, 2013a), and measured soil inorganic carbon (SIC) up to 1 meter (695–1738 Pg) (Eswaran et al., 2000; Hirmas et al., 2010); (4) the atmosphere (863 Pg) and (5) the biosphere (540–610 Pg) (Rice, 2004).

Abbreviations: MAP, mean annual precipitation; MAT, mean annual temperature; MAI, mean annual aridity index; MAPET, mean annual potential evapotranspiration; AWB, annual water balance; WSWB, wet season water balance; BD, bulk density; SOC, soil organic carbon; SIC, soil inorganic carbon; Nt, total nitrogen; C/N, carbon/nitrogen ratio; SOC_s, SOC storage; SIC_s, SIC storage; STC_s, STC storage; CF, coarse fragments.

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Many studies have been concentrated on the changes of STC in the topsoil (i.e., 0–20 cm depth) (Harrison et al., 2011) due to the ease of sampling and data collection, but there have been few studies on STC changes in the deeper soil horizon. However, the subsurface soil has a large carbon storage capacity (Jobbágy and Jackson, 2000), and there is a lot of evidence showing that the STC content of subsoil is sensitive to climate changes, land use (Knops and Kate, 2009; Carter and Gregorich, 2010) and management (Khan et al., 2007).

The soil C pool consists of two distinct components: SOC and SIC (Zhang et al., 2015). The SOC component, a key indicator of soil quality, influences the essential biological, chemical and physical soil functions such as nutrient cycling, water retention and soil-structure maintenance (Vitti et al., 2016). Due to the sensitivity to environmental changes, SOC is one of the most important components involved in global climate change (Lal, 2004; Selim et al., 2016).

Climate and parent materials can introduce a range of C levels in the ecosystems (McLauchlan, 2006). Jenny (1980) expressed that the best climosequence was observed over long transects with gradual slope gradients. Climate provides water and temperature (energy), two main components involved in soil formation. Changes in precipitation and temperature are responsible for the biomass entering to soil, which can also manipulate soil properties. Increasing rainfall decreases soil pH, while increasing the maximum depth of carbonate accumulation, as well as SOC, nitrogen and clay contents (Buol et al., 2011).

Soil formation needs long time sequences and it is subjected to different climates. Furthermore, chemical, physical and mineralogical

properties of parent rocks play a key role in soil formation. Soils formed on ferromagnesian rocks, such as basalt, are usually rich in nutrients such as iron, manganese, potassium, calcium, magnesium and phosphorus. Basalt weathers rapidly and yields fine textured (clayey) soils (Buol et al., 2011).

SIC (both lithogenic (LIC) and pedogenic (PIC)) (Batjes, 1996; Zamanian et al., 2016) also plays a significant role in carbon sequestration (Chang et al., 2012; Monger et al., 2015). Inorganic carbon is not considered as a carbon stock in several studies (Wu et al., 2009; Tan et al., 2014), probably due to the longer time needed for changes in carbonates compared to the shorter time for SOC (Rawlins et al., 2011; Yang et al., 2012). Recent studies, however, have demonstrated that soil biota plays an important role in the generation of secondary carbonates (Lee et al., 2008), highlighting potentially rapid changes in SIC. Monger et al. (2015) suggested that the source of calcium (Ca) is the main factor in CO₂ sequestration by soil inorganic carbon. Pedogenic calcium carbonate (CaCO₃) resulting from limestone weathering, however, does not sequester atmospheric carbon because the source of the Ca is from preexisting CaCO₃, and the CO₂ consumed in the reaction to form carbonic acid is released upon the reprecipitation of CaCO₃ (Drees et al., 2001). In contrast, if the Ca or Mg derives from the weathering of silicates bearing Ca-Mg, two moles of CO₂ is consumed per mole of carbonate precipitated, and CO₂ is sequestered during this process.

Yang et al. (2012) stated that acidification leads to a significant decrease in carbonates in some Chinese soils, especially, those that contained higher amounts of inorganic carbon. This study proves that SIC is dynamic rather than the conventional concept considering it a stable C pool. Sedimentation of carbonates and/or release of CO₂ to the atmosphere from SIC acidification influence the terrestrial carbon sequestration processes. SIC migration from the surface horizon to the deeper ones or down slopes in landscapes are other important processes leading to the decrease of SIC in soils. The emitted CO₂ from SIC sources may influence the terrestrial C cycle in the ecosystems resulting in global warming. Without considering SIC dynamics, an understanding of the C cycle remains incomplete; thus SIC should be regarded as one of the C stocks (Yang et al., 2012).

SIC pools and dynamics in the arid and semi-arid regions, which cover around one-third of the earth's surface, are very important due to their overall higher accumulation rate of SIC than the accumulation seen in other biomes (Drees et al., 2001; Lal, 2004, 2008). The greatest extent of the SIC stock is found under arid, semi-arid and Mediterranean conditions with 77.8%, 14.2% and 5.4%, respectively (Eswaran et al., 2000; Jobbágy and Jackson, 2000). There is an equilibrium between the soil components and the environment; therefore, SOC and SIC, which originated from the same sources during soil formation, reflect the existing thermodynamic equilibrium across the period of soil evolution. With the above-mentioned considerations, the current study compares organic and inorganic carbon content and storage in natural soil pedons along an arid to dry sub-humid climosequence in the soils of northwestern Iran, aiming to: 1) evaluating the relationship and variability of SOC and/or SIC content and some soil properties between the solum and the underlying rock; and (2) quantifying the soil organic and inorganic carbon storage along an arid to dry sub-humid climosequence.

2. Materials and methods

2.1. Study area and field sampling

The study was conducted at three regions in the arid, semi-arid and dry sub-humid areas as a climosequence located in northwestern Iran (Fig. 1). Stable positions without anthropic effects were selected to reduce the effect of other variables in the relationships between SOC and SIC.

The geographic midpoint coordinates of the studied regions are shown in Table 1. Soil moisture and temperature regimes and evapotranspiration were determined using long-term mean annual precipitation and temperature data by jNSM software (java Newhall Simulation Model) (USDA-NRCS, 2012b). Aridity index (AI) = MAP / MAEP (UNEP, 1997) was used to quantify precipitation availability over atmospheric water demand (Table 1).

The studied areas located in the Alborz geological zone. Bedrock underlying the soil in all areas consisted of mid to late Eocene basalts (Sahandi and Soheili, 2005). After identification of the areas on geology maps (1:100,000), parent rocks were checked by field control. The arid region consisted of very thick, coarse, textured, massive, non-porous and homogenous megaporphyric trachybasalt without fracture (Chaichi and Mohabbi, 1995). The semi-arid region composed of very thick, textured, massive structure with some planar cracks and homogenous gray basalt, trachybasalt and andesitic basalt complex (Radfar, 2002). The dry sub-humid region contained very thick, fine, textured, massive (with some veins and pores in fractures) and homogenous basaltic lava (Nazari and Salamati, 1998). The studied regions had hilly geomorphic units and about 220–2200 m elevation above sea level. No conflicts with tephra episodes were observed for the basalts in the regions. There was no evidence of covering layers on basalts before denudation; thus, different soil thicknesses in the regions were the results of weathering of underlying rocks (Fig. 2). The upper boundaries of basalt were limited to soil thicknesses in each region (40–80 cm in arid, 80–120 cm in semi-arid and 150 to >250 cm in dry sub-humid), but their lower depths went beyond 100 m up to 200 m in different regions (Sahandi and Soheili, 2005). Depending on climatic conditions, weathering intensity in different regions varied between low in the arid region to high in the semihumid region. Field observations, physical characteristics and the results of clay mineralogy of surface and subsurface horizons (data not shown) did not show any evidence of dust deposition. All the sampling sites were on stable upland positions with a southeast to northwest aspect and slopes averaging less than 25%. The sampled pedons did not exhibit evidence of human perturbation or accelerated wind and water sedimentation.

Land use in the studied regions was rangeland with somewhat different vegetation cover in each region in accordance with their climatological properties. The vegetation cover in the arid region consisted of mainly annual herbs, along with sparse perennial shrubs with low fine-root density. However, in the semi-arid and dry sub-humid regions more intensive deep-rooted annual grasses or herbs, accompanied by perennial shrubs and small trees with moderate to high-root densities, were observed. Ten pedons inside hill geomorphic units from each region (about 500 ha extent each, 30 pedons in 1500 ha in total) were described and sampled according to the standard methods (USDA-NRCS, 2012a). The soils were classified based on Keys to Soil Taxonomy (Soil Survey Staff, 2014).

2.2. Physico-chemical analyses

27 out of 30 pedons comprising 85 samples from the genetic horizons/layers were selected for laboratory analyses. After complete analyses of the samples and classifying the soils according to Keys to Soil Taxonomy (Soil Survey Staff, 2014), nine bulk pedons with the most distinctive characteristics of the soil taxonomic units were selected to be presented (Table 2). All analyses were performed on air-dried and sieved (2 mm sieve) soil samples (Pansu and Gautheyrou, 2006). Coarse fragments' percentage was determined from the weight of fragments with diameter >2 mm/weight of whole sample ratio multiplied by 100 (SCS-USDA, 1967). Particle size distribution in fine earth fractions was determined using the hydrometer method (Gee and Bauder, 1986). Bulk density was determined by the core method (Blake and Hartge, 1986). SOC and SIC (as calcium carbonate equivalent (CCE)) content was determined using the Walkley-Black and calcimetry methods (with increasing reaction time), respectively (Carter and

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