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Afforestation effects on soil organic carbon and nitrogen pools modulated by lithology

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ABSTRACT

Afforestation is regarded as an effective way to sequester soil organic carbon (C) and hence to attenuate global warming. Nevertheless, so far mechanisms underlying the direction and magnitude of soil C pool variation following afforestation are not well understood. Here soil organic C, total nitrogen (N) and total phosphorus (P) pools in surface mineral soil (0–15 cm) were measured in cropland and plantation forest over two lithology types, i.e., limestone and clasolite, using a paired-site approach in a region of southwest China. The major objective was to test whether the effects of afforestation on soil C, N and P pools were affected by lithology. We found that lithology modulated the effect of afforestation on soil C or N pool, but not on soil P pool. Soil C pool was not altered by afforestation over limestone, but was significantly increased over clasolite ($P < 0.001$). Instead, soil N pool was significantly decreased by afforestation over limestone ($P < 0.001$), but was not altered over clasolite. Soil P pool was consistently decreased by afforestation regardless of lithology type ($P < 0.05$). The molar ratios among soil C, N and P were not affected by lithology, but were significantly elevated by afforestation ($P < 0.01$), indicating that soil C sequestration would increasingly be limited by soil N or P following afforestation. Our study suggests that afforestation effects on soil C and N pools may be controlled by lithology, and thus provides a new mechanism to explain the divergent responses of soil C and N following afforestation.

1. Introduction

Soil organic matter (SOM) is the largest reservoir of carbon (C) in active exchange with atmospheric carbon dioxide, and thus the changes of soil organic C (soil C hereafter) pool exert substantial impacts on global climate (Torn et al., 2009). Afforestation can increase sequestration of atmospheric carbon dioxide and hence attenuate global warming (Bonan, 2008). Nevertheless, uncertainty still exists in predicting the direction and magnitude of soil C pool change upon afforestation, largely due to poor understanding of the mechanisms underlying soil C pool change following afforestation. Meta-analyses generally show that afforestation increases soil C pool, but the patterns vary according to sites or regions (Guo and Gifford, 2002; Don et al., 2011; Li et al., 2012; Deng et al., 2014; Wei et al., 2014). For example, soil C pools have been reported to increase, decrease or change negligibly after afforestation in case studies (Li et al., 2012; Deng et al., 2014). A few factors such as climate, prior land use, tree species planted and soil type have been suggested to explain the differential afforestation effects (Guo and Gifford, 2002; Don et al., 2011; Li et al., 2012; Assad et al., 2013; Deng et al., 2014). However, in many cases, afforestation with the same species on similar prior land use types under the same climate resulted in difference in both direction and magnitude of soil C pool change according to a meta-analysis (Li et al., 2012). This clearly demonstrates that some other factors which are crucial in determining soil C pool change following afforestation may have been overlooked.

Surficial lithology, i.e., the geochemical, mineralogical, and physical properties of rocks (Hartmann and Moosdorf, 2012), is among the overlooked factors which has the potential to strongly affect soil C pool. The lithology of parent materials can greatly affect both the chemical and physical properties of soils (Anderson, 1988; Kosmas et al., 2000; Leuschner et al., 2006; Neff et al., 2006; Gray et al., 2014). This in turn has a variety of influences on soil C pool via its effects on plant productivity (C inputs to soil), soil C decomposition and stabilization (Kooijman et al., 2005; Torn et al., 2009; Aranda et al., 2011). Studies that

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relate soil C to lithology show that lithology may have or not have significant effects on soil C pools (Garcia-Pausas et al., 2007; Oyonarte et al., 2008; Turrión et al., 2009; Fantappiè et al., 2010; Albaladejo et al., 2012). For example, soil C pool was significantly higher in soils underlain by schist than by granite (Turrión et al., 2009). Nevertheless, another study showed that soil C level was not significantly different, although some soil properties differed between soils underlain by calcareous and metamorphic materials (Oyonarte et al., 2008). Based on a study along a 4000-km-long north–south transect of natural grassland and shrubland in Chile and the Antarctic Peninsula, Doetterl et al. (2015) reported that the interactions of climatic and geochemical factors controlled soil C pool and turnover. This indirectly supports the crucial role of lithology in soil C pool change as geochemical factors are largely controlled by lithology (Gray et al., 2014). In addition, inclusion of lithology in models was found to greatly improve the model performance in predicting soil C distribution in eastern Australia (Gray et al., 2014), which also suggests that lithology is crucial in determining soil C pool change.

Although accumulating evidence indicates that lithology is a major regulating factor on soil C level, whether or not lithology controls the direction and magnitude of soil C pool change following afforestation, to our knowledge, has rarely been reported (Kosmas et al., 2000). Since lithology has a variety of influences on soil physicochemical properties, theoretically it has the potential to affect soil C pool following afforestation. This is supported by Kosmas et al. (2000), who reported that SOM increased greatly over volcanic lava and schist-marble, but not over shale following agricultural abandonment in Greece. In order to test whether the effect of lithology on direction and magnitude of soil C pool change after agricultural abandonment including afforestation is widespread, more studies are undoubtedly needed.

The paired-site approach is widely used to explore the effects of afforestation on soil C pool change (Paul et al., 2002; Laganiere et al., 2010; Li et al., 2012). In the current study, soil C pools in corn-soybean field and pine (*Pinus massoniana*) plantation of the first rotation over two contrasting lithology types, i.e., limestone and clasolite, were investigated in a region of southwest China. The major objective was to test whether the effects of afforestation on soil C pool were modulated by lithology. As there is evidence showing that the effects of land use change on soil C pool change depend on lithology type (Kosmas et al., 2000), we hypothesized that afforestation would differentially affect soil C pools over these two lithology types due to their substantial difference in soil properties (Cao et al., 2011). Furthermore, soil N and total phosphorus (soil P hereafter) are the two major limiting nutrients for plant productivity, and hence are crucial in determining soil C accumulation following afforestation. In this regard, we measured soil N and P as well in order to assess the possible limitation of plant productivity and soil C sequestration by soil N and P following afforestation. Since soil C, N and P are usually closely coupled (Finzi et al., 2011), we hypothesized that they would change in similar patterns after afforestation over both lithology types.

2. Methods

2.1. Study region

The study region (23°40′N–25°25′N, 107°35′E–108°30′E) is located in the northwest of Guangxi Zhuang Autonomous Region, southwest China. Mean annual air temperature is 17.8–22.2 °C, with the lowest monthly mean in January (7.8–13.0 °C) and the highest in July (25.8–29.4 °C). Mean annual precipitation ranges from 1346 to 1640 mm with a distinct seasonal pattern. The period from April to September is the wet season and that from October to March is the dry season. The region is mountainous and interwoven with karst areas and non-karst areas. Both types of areas are characterized by gentle valleys flanked by hills. The lithology in the karst areas is mostly limestone and, to a lesser extent, dolomite and their mixtures, while the lithology in non-karst areas is clasolite. The soil is calcareous lithosols (limestone soil) over limestone, and is ferralsols over clasolite according to the FAO/UNESCO classification system. Although the study region is located within the subtropical evergreen broadleaf forest zone, plantation forest, especially pine (*Pinus massoniana*) plantations are widely distributed.

2.2. Soil sampling and analyses

A paired-site approach was adopted in this study (Fig. S1). The sampling covered areas over limestone and clasolite. For each lithology type, five pairs of toposequences were identified. Each pair of toposequences included a corn-soybean (or corn field hereafter) rotation cropland toposequence and a pine plantation toposequence with similar slope and aspect. For each toposequence of cropland or plantation, three sampling locations in the valley, foot slope and back slope, respectively, were chosen. The slope was typically 15 to 20 degrees and the aspect was south to southeast facing. The distance between any two sampling locations within a pair of toposequences was normally less than 20 km except an extreme case where the distance was about 60 km. The plantations were limited to those of the first rotation with an age of about 30–40 years and the prior land use being corn-soybean field as well. The history of the land use was obtained by inquiring native people. Therefore, uncertainties existed in terms of age of plantation and prior land use management. The understory species included *Dicranopteris pedata*, Rhodomyrtus tomentosa, *Blechnum orientale*, *Melastoma dodecandrum*, *Arthraxon hispidus*. The croplands were managed under corn-soybean rotation each year, and fertilized with compound fertilizer with N, P and K fertilization rates of about 150, 60 and 120 kg ha^{-1} yr^{-1}. Tillage was usually not adopted, but disturbance associated with planting and weeding to the surface soil was unavoidable. In total, 60 sampling locations (i.e., 5 paired toposequences × 2 land use types × 3 sampling locations × 2 lithology types) were selected.

Soil sampling was conducted from the end of March to early June 2015. At each location, a plot of about 20 m × 20 m was selected. For most of the sampling sites, obvious organic layer was absent, so the organic layer was not collected. Since soil depth was much heterogeneous and only shallow soil layers could be found in most of the sampling sites over limestone, mineral soils of 0–15 cm depth were collected after removal of organic layer (if available) in order to make comparison among sampling sites. Ten to fifteen soil cores (0–15 cm in depth and 5 cm in diameter) were randomly collected for each plot and mixed to a composite sample. Additional soil cores were collected to determine bulk density (BD). Soils were air-dried and sieved through a 2-mm mesh sieve to remove roots and stone fragments.

Soil organic carbon was measured by wet oxidation with dichromate redox colorimetric method, with which carbonates are not determined (Carter and Gregorich, 2006). Soil pH (1:2.5 soil/water ratio) was measured with a pH meter (FE20K, Mettler-Toledo, Switzerland). Soil N was analyzed using an elemental analyzer (EA 3000; EuroVector, Italy). Soil total P was determined by acid digestion with a $H_2SO_4 + HClO_4$ solution (Carter and Gregorich, 2006). Soil texture was determined using a laser diffraction particle size analyzer (Mastersizer, 2000, Malvern, UK). Exchangeable calcium (Ca), magnesium (Mg), potassium (K) and sodium (Na) were displaced via compulsive exchange in 1 mol L^{-1} ammonium acetate at pH 7.0 and analyzed by inductively coupled plasma atomic emission spectroscopy (ICP-AES) (Hendershot et al.,