Available online at www.sciencedirect.com
ScienceDirect
Journal of Environmental Sciences 19(2007) 319–326
JOURNAL OF ENVIRONMENTAL SCIENCES
ISSN 1001-0742
CN 11-2629/X
www.jesc.ac.cn

Topsoil organic carbon mineralization and CO_2 evolution of three paddy soils from South China and the temperature dependence

ZHANG Xu-hui, LI Lian-qing, PAN Gen-xing*

Institute of Resources, Ecosystem and Environment of Agriculture, Nanjing Agricultural University, Nanjing 210095, China.
E-mail: xuhuizhang@njau.edu.cn

Received 5 April 2006; revised 15 June 2006; accepted 21 July 2006

Abstract

Carbon mineralization and its response to climatic warming have been receiving global attention for the last decade. Although the virtual influence of temperature effect is still in great debate, little is known on the mineralization of organic carbon (SOC) of paddy soils of China under warming. SOC mineralization of three major types of China's paddy soils is studied through laboratory incubation for 114 d under soil moisture regime of 70% water holding capacity at 20°C and 25°C respectively. The carbon that mineralized as CO_2 evolved was measured every day in the first 32 d and every two days in the following days. Carbon mineralized during the 114 d incubation ranged from 3.51 to 9.22 mg CO_2-C/gC at 20°C and from 4.24 to 11.35 mg CO_2-C/gC at 25°C respectively; and a mineralizable C pool in the range of 0.24 to 0.59 gC/kg, varying with different soils. The whole course of C mineralization in the 114 d incubation could be divided into three stages of varying rates, representing the three subpools of the total mineralizable C: very actively mineralized C at 1–23 d, actively mineralized C at 24–74 d and a slowly mineralized pool with low and more or less stabilized C mineralization rate at 75–114 d. The calculated Q_{10} values ranged from 1.0 to 2.4, varying with the soil types and N status. Neither the total SOC pool nor the labile C pool could account for the total mineralization potential of the soils studied, despite a well correlation of labile C with the shortly and actively mineralized C, which were shown in sensitive response to soil warming. However, the portion of microbial C pool and the soil C/N ratio controlled the C mineralization and the temperature dependence. Therefore, C sequestration may not result in an increase of C mineralization proportionally. The relative control of C bioavailability and microbial metabolic activity on C mineralization with respect to stabilization of sequestered C in the paddy soils of China is to be further studied.

Key words: C mineralization; carbon pool; laboratory incubation; paddy soil; soil warming

Introduction

Soil carbon mineralization and the CO_2 evolution has been paid great attention for its important effect on the global carbon cycle and terrestrial ecosystem functioning (IPCC, 2001; Valentini *et al*., 2000; Jenkinson *et al*., 1991). The global mean air temperature is predicted to increase 1.0–3.5°C by the year 2100 (Houghton *et al*., 1995, 1996). Soil organic carbon (SOC) dynamics and the impact of SOC mineralization on CO_2 evolution in response to global warming have been in vigorous debate. Although Giardina and Ryan (2000) argue that soil warming do not necessarily affect SOC mineralization, there have been numerous studies dealing with SOC mineralization under simulated warming and its potential impact on the soil-air CO_2 flux. Kirschbaum (1995, 2000) has analyzed many references of SOC mineralization using soil incubation and inferred that the mineralization of SOC is correlated to soil temperature, indicating that soil warming impacts the soil dynamics and soil-air CO_2 flux. Oechel *et al*. (1993, 1995), have evidenced that the mass loss of the soil organic carbon from the habitat of the tundra in Alaska has resulted from the climatic warming. Goulden *et al*. (1998), have found that there has been a loss of soil carbon in the boreal forest, in the early spring, in Manitoba, Canada, because of warming. By employing laboratory incubations of soil columns of whole soil or of various soil layers, Bowden *et al*. (1998), Macdonald *et al*. (1999), and Leirs *et al*. (1999), studied the decomposition of the soil organic matter of forest soils and they concluded that C mineralization and the temperature dependence of forest soils varied with soil types, horizons, soil N availability, soil moisture regimes, and so on.

Nevertheless, few studies have been reported on C mineralization in the agricultural soils. Paddy soil is considered to be a unique anthropogenic soil in China with an area of 30 Mhm^2, and one quarter of the China's national cereals production (Pan *et al*., 2004; SSSSC, 1998). Pan *et al*. (2003, 2004), has reported that the China paddy soils have a greater potential of C sequestration than the upland cultivated soils and a remarkable tendency of C enrichments has occurred in these paddy soils in the last two decades. The winter warming, as a result of global

Project supportrd by the National Natural Science Foundation of China (No. 40231016, 40171052). *Corresponding author.
E-mail: pangenxing@yahoo.com.cn.

climatic change, has been observed recently in Southern China, where 90% of the China paddy soils are distributed (Gao *et al*., 2003; Li, 1992). Therefore, the mineralization of paddy SOC and its potential response to global warming may be of great concern to the C dynamics of agricultural soil in the context of global climate change The purpose of this study is to examine the mineralization of SOC from the different types of paddy soils of China, using laboratory incubation, with special reference to the soil warming effects on C mineralization and CO_2 evolution. The authors also aim at clarifying whether or not C sequestration in paddy soils will result in the proportional increase of C mineralization and the CO_2 emission potential.

1 Materials and methods

1.1 Soils used

Three paddy soils, representing the predominant types of paddy soils of South China, were chosen for this study. These soils were Soil PP from Beipei, Chongqing, Soil RP from Jinxian, Jiangxi, and Soil EP from Xiantao, Hubei. Their classification and genesis are shown in Table 1. These soils have been cultivated with rice for more than 50 years and the present rotation is mainly rice and rape. The topsoil samples of 0–15 cm depth have all been sampled in January 2003. After shipping to the laboratory, the samples were air-dried and ground to pass through a 2-mm sieve.

1.2 Experiment design

In this study, the mineralization experiment was conducted using controlled laboratory incubation. PVC cylinders were used for incubation with a diameter of 7.80 cm and a height of 25 cm and the incubation design was as shown in Fig.1. Ground soil was put into a cylinder to an approximate height of 15 cm and a specific volumetric weight of 1.2 g/cm^3. The cylinders were then put in a beaker with a water table close to the bottom of the soil, so as to keep wet soil moisture simulating the field condition of paddy (approximately 70% of WHC, and the measured water content (%) during the incubation of Soil PP, RP, and EP was 44.9±1.9, 57.3±3.0 and 44.0±0.8 respectively). The moisture maintenance in the soil cylinders was done by weight balance every day. The incubation was conducted for 114 d in two LRH-250-S temperature and humidity incubators adjusted to 20 and 25°C respectively. The C mineralization was determined as CO_2 evolved (sampled at 10 p.m. every day in the first 40 d and every two days thereafter till the end). The experiment was done in triplicates.

1.3 Basic soil properties

Determination of basic properties of the soils was done following the conventional laboratory methods described by Lu (2000). Soil pH was measured by using a Metter-Toledo pH meter with a soil:water ratio of 1:2.5, cation exchange capacity (CEC) by pH 7, 1 mol/L ammonium acetate, clay (< 0.002 mm) content with a hydrometer, and free iron oxyhydrates (Fe_d) with dithionate-citrate-bicarbonate extractions respectively. Total organic carbon (TOC) and total nitrogen (TN) of dried samples were measured after decarbonization with HCl (10% v/v) using a CNS macroelemental analyzer (Vario-Max Elementar Analysensysteme GmbH, Germany). Moisture of the samples was determined by oven-drying at 105°C for 6 h. The results of the determinations are listed in Table 2.

1.4 CO_2 evolution

Collection and analysis of CO_2 evolved was conducted by following a procedure described by Coleman *et al*. (1978). CO_2 evolved was absorbed in a 50-ml beaker containing 25 ml of 0.1 mol/L NaOH solution. Total CO_2-C trapped in the alkaline solution was precipitated as

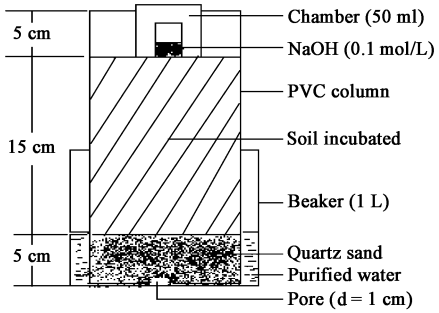

Fig. 1 Soil column design used for incubation.

Table 1 Paddy soils used in this study

Soil	US soil taxonomy[1]	Chinese soil classification[2]	Chinese soil taxonomy[3]	GPS position	Parent material
PP	Typical Haplaquept	Purple soil	Haphydroagric Anthrosol	N: 29°48.597′ E:106°25.584′	Weathered purple sandstone
RP	Typical Hapludult	Typical red soil	Ultic Hydroagric Anthrosol	N: 28°21.133′ E:116°10.551′	Quaternary weathered deposit
EP	Typical Fluvaquent	Gray meadow soil	Entic Hydroagric Anthrosol	N: 30°22.477′ E:113°21.286′	Riverine alluvium

1: US soil taxonomy (Soil Survey Staff of USDA, 1999); 2: Chinese soil classification (SSSSC, 1998); 3: Chinese soil taxonomy (Gong, 1999).

Table 2 Soil basic properties of the paddy soils studied

Soil	pH (H_2O)	SOC (g/kg)	TN (g/kg)	CEC (cmol(+)/kg)	Fe_d (g Fe/kg)	Clay (g/kg)	C/N
PP	6.89	17.48	1.35	26.02	13.46	265.0	12.99
RP	5.54	24.39	2.23	9.38	23.38	259.0	10.94
EP	7.15	17.71	1.76	25.38	21.97	453.5	10.05

barium carbonate with $BaCl_2$ and determined by titration of the excess NaOH to pH 8.3 with 0.1 mol/L HCl.

1.5 C pool analysis

Soil particulate organic carbon (POC) was determined by following a procedure described by Cambardella and Elliott (1992): Twenty grams of soil passed through a 2-mm sieve was shaken in 100 ml 0.5 mol/L NaOH for 18 h, then the suspension was passed through a 53-μm sieve. The remnants on the sieve were the particulate material, in a size larger than 53 μm, and these were washed with distilled water. This material was then transferred to a dried bottle and weighed after drying at 55°C for 72 h. Sub-samples of bulk soil and of the particulate material fraction were ground and analyzed for total C and N, using dry combustion at 900°C with a CNS elemental analyzer (Elementar Analysensysteme GmbH, Germany).

Labile carbon pool (LC) was determined by a procedure recommended by Blair *et al*. (1995). The organic carbon oxidized by 0.333 mol/L $KMnO_4$ was defined as labile carbon. The procedure was as follows: Samples of soil containing about 15 mg C were weighed into 50 ml plastic screw top centrifuge tubes, and 25 ml of 0.333 mol/L $KMnO_4$ was added to each vial. Blank samples, containing no soil, and samples of a standard soil were analyzed in each run. The centrifuge tubes were tightly sealed and tumbled for 1 h , at 12 r/min, in a tumbler with a radius of 15 cm. The tubes were centrifuged for 5 min at 2000 r/min (RCF=815 g) and the supernatants were diluted to 1:250 with deionized water. The absorbencies of the diluted samples and standards were read on a split beam spectrophotometer at 565 nm. The range of the standards was chosen to adequately cover the sample range, normally 300 to 333 mol/L. The change in the concentration of $KMnO_4$ was used to estimate the amount of the carbon that was oxidized, assuming that 1 mmol/L MnO_4 was consumed in the oxidation of 0.75 mmol/L, or 9 mg of carbon. The results were expressed as gC/kg soil.

Soil microbial biomass carbon (MBC) was determined using the chloroform fumigation-extraction method described by Jenkinson and Powlson (1976). Fresh soil of 25 g, was fumigated at 25°C for 24 h prior to extraction by K_2SO_4. Fumigated and unfumigated samples were extracted respectively with 100 ml of 0.5 mol/L K_2SO_4 for 1 h. The organic C in the extracts was measured by oxidation with dichromate, and the MBC was expressed as the difference in C between the unfumigated and fumigated samples by following Voroney *et al*. (1993).

Dissolved organic carbon (DOC) was determined by extracting an unfumigated soil sample with 0.5 mol/L K_2SO_4 solution (Joergensen *et al*., 1996). The extract was analyzed for soluble organic C using the potassium dichromate method (Jenkinson and Powlson, 1976).

The results of the original C pools of the studied soils are given in Table 3.

1.6 $\delta^{13}C$ analysis of the CO_2 evolved

The $\delta^{13}C$ of mineralized C of the evolved CO_2, collected on selected days during incubation, was analyzed. The evolved CO_2 was trapped in 0.1 mol/L NaOH solution and precipitated with saturated $BaCl_2$ as $BaCO_3$. The $\delta^{13}C$ of the collected $BaCO_3$ precipitates was determined using Mass Spectrometry (MAT 252, Finnigan Co. Ltd, UK), at the isotope geochemistry lab of the State Key Lab of Environmental Geochemistry, Guiyang, China.

Table 3 C pools of the studied soils prior to incubation

Soil	TOC (g/kg)	POC (g/kg)	Labile C (g/kg)	MBC (mg/kg)	DOC (mg/kg)
PP	17.48	4.47	6.76	635.07	45.59
RP	24.39	3.84	10.56	961.52	98.23
EP	17.71	2.56	7.40	871.33	87.98

2 Results

2.1 Total mineralization and CO_2 evolution

The total mineralized C and CO_2 evolved from different soils, both for different soils under different temperature treatments for 114 d incubation is presented in Table 4. During 114 d of incubation, total mineralized C ranged from 228 to 330 mg under 20°C and from 288 to 427 mg under 25°C. The mean C mineralization rate varied in the range of 0.13 to 0.20 mg CO_2-C/(g OC·d), also being significantly higher under higher temperature. The C mineralized in the whole experiment for 114 d could be considered as the potentially mineralizable C pool of these soils, which ranged from 0.13 to 0.16 mgC/(gOC·d) at 20°C and from 0.17 to 0.20 mgC/(gOC·d) at 25°C respectively.

2.2 Dynamics of C mineralization and CO_2 evolution

The daily CO_2 evolution from C mineralization was plotted as shown in Fig.2. Generally, C mineralization underwent an increasing stage after the initiation of the incubation followed by a decreasing one after 23 d under both temperatures. C mineralization reached a peak in a time course of about three weeks. Later on, C mineralization decreased consistently with time. However, C mineralization in 24–74 d decreased more sharply than in the last stage of incubation, between 75–114 d.

C mineralized and the mineralization rate was grouped in Table 5. Apparently, C mineralization rate varied with the different stages, being highest in the first stage. Widest variation of C mineralization rate with different soils was found in the first stage, being almost double than that in the last stage. It seemed that the mineralizable C pool left after 74 d of incubation was increasingly inaccessible to microbial attack in all the soils. The mean percentage of C mineralized in one stage to that of the total is 25%, 50%, and 25% for the first, second, and third stages respectively. Soil warming effect on C mineralization also varied with different stages. Significant difference in C mineralization rate under two temperatures could be detectable in all stages for soil EP only. Thus, the warming effect tended to depend on the soil type and/or the length of soil incubation in the laboratory study.

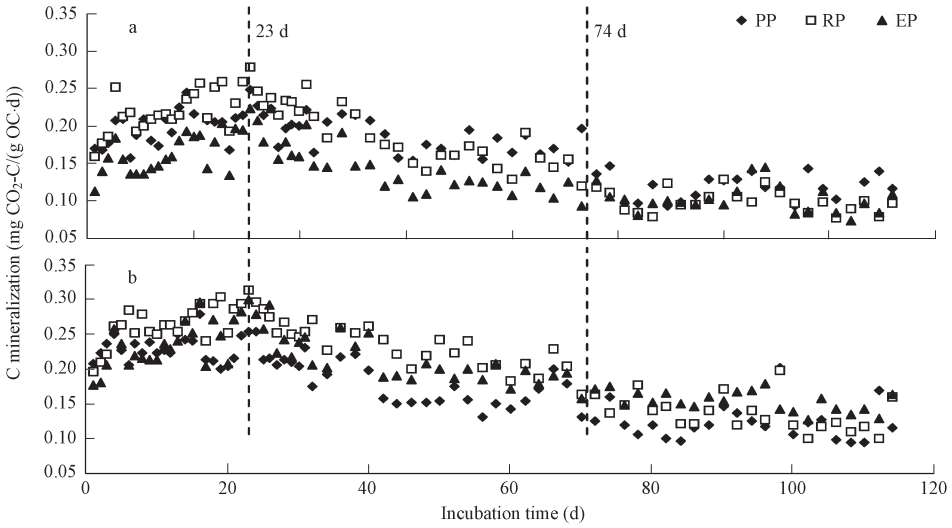

Fig. 2 Dynamics of C mineralization as CO_2 evolution during the incubation under 20°C(a) and 25°C(b).

Table 4 C mineralization as CO_2 evolved during the incubation and the warming effect

Soil	Total CO_2-C evolved (mg CO_2-C)		Mineralization rate (mgC/(g OC·d))		Q_{10} value			
	20°C	25°C	20°C	25°C	Whole	1–23 d	24–74 d	75–114 d
PP	283.52±18.48Ba	288.32±18.48Aa	0.16±0.01Ca	0.17±0.01Aa	1.04	1.30	0.92	1.05
RP	330.03±43.11Ca	427.09±42.46Cb	0.15±0.02Ba	0.20±0.02Cb	1.67	1.43	1.65	2.09
EP	227.75±15.51Aa	344.03±9.58Bb	0.13±0.01Aa	0.19±0.01Bb	2.27	2.06	2.29	2.42

Different capital and lower case letters indicate significant difference between the different soils and different temperatures respectively at $p<0.05$.

Table 5 Variation of C mineralization in the different stages of incubation*

Soil incubated		Total C mineralized (mgC/g OC)			Mean C mineralization rate (mgC/(g OC·d))		
		1–23 d	24–74 d	75–114 d	1–23 d	24–74 d	75–114 d
PP	20°C	4.64±0.18 (24.94)	9.22±0.24 (49.55)	4.75±0.73 (25.50)	0.20±0.01	0.18±0.01	0.12±0.02
	25°C	5.29±0.57 (27.81)	8.86±0.60 (46.62)	4.86±0.14 (25.57)	0.23±0.03	0.17±0.01	0.12±0.00
RP	20°C	5.08±0.18 (28.79)	8.84±1.32 (50.09)	3.73±0.95 (21.12)	0.22±0.01	0.17±0.03	0.10±0.02
	25°C	6.09±0.36 (26.66)	11.35±1.24 (49.72)	5.39±0.75 (23.63)	0.27±0.02	0.22±0.02	0.14±0.02
EP	20°C	3.78±0.31 (25.70)	6.91±0.78 (46.92)	4.03±0.12 (27.38)	0.16±0.01	0.14±0.02	0.10±0.00
	25°C	5.42±0.09 (24.49)	10.46±0.36 (47.22)	6.27±0.33 (28.29)	0.24±0.00	0.21±0.01	0.16±0.03

*The number in the parenthesis represents the percentage of C mineralized in each stage to the total C mineralized after 114 d of incubation.

2.3 Change of $\delta^{13}C$ of mineralization-evolved C

The results of the selected samples of the evolved CO_2 upon C mineralization are listed in Table 6. The initial $\delta^{13}C$ values at the beginning of incubation of the soils were similar (–22.35±1.00 ‰ PDB), indicating a dominant C-3 plant origin of soil organic matter in paddy. Li *et al.* (2000), reported the $\delta^{13}C$ of topsoil SOM of two paddy soils from southern Jiangsu that ranged from –28.7 ‰ PDB to –22.8‰ PDB. There was a general tendency of an increase in the $\delta^{13}C$ value of the evolved CO_2, with the time of incubation for PP and RP, whereas the $\delta^{13}C$ of the evolved CO_2 from the soil RP and EP at the first two stages appeared to be slightly heavier under warming. This could evidence that the more stable C was attacked as active C pool during the 1–23 d incubation. The $\delta^{13}C$ value of the evolved CO_2 from soil PP and RP at the end of incubation was significantly heavier than that at 1 d or at 24 d of incubation, demonstrating the C mineralized in the last stage originated from a more stable C pool than at the end of the first stage. This again supported the idea that the decreased C mineralization resulted from the attack of a C pool of increased stability. However, the trend of $\delta^{13}C$ values of the evolved CO_2 of the soils at 114 d, was contrary to that of the original trend at 1 d. This could imply a different utilization of C source for the microbial isotopic fractionation during decomposition and CO_2 evolution under incubation, which deserves further study.

Table 6 $\delta^{13}C$ (‰ PDB) of the evolved CO_2 sampled at different incubation days under 20 and 25°C

Soil	1 d		24 d		114 d	
	20°C	25°C	20°C	25°C	20°C	25°C
PP	–23.42	ND	ND	–23.65	–18.32	–20.85
RP	–22.84	–22.78	–21.45	–21.31	–20.14	–21.78
EP	–22.03	–21.98	–24.00	–23.98	–23.08	–23.09

ND: not detected.

3 Discussion

3.1 C mineralization and the temperature dependence of the paddy soils

Many studies have been reported on C mineralization

and CO_2 evolution of natural and cropland soil, both under treated and untreated conditions (Collins *et al*., 2000; Henriksen and Breland, 1999; Reichstein *et al*., 2000; Schimel *et al*., 1999; Kruse *et al*., 2004). Côté *et al*. (2000), studied C mineralization of forest topsoil by incubation at constant 21°C for 282 d, and found that the C mineralization of the mineral topsoil ranged from 0.12 to 0.45 mgC/(g OC·d). Reichstein *et al*. (2000), conducted an experiment of C mineralization of forest soils under 5, 15, and 25°C with 60% of WHC condition and reported a mineralization rate ranging from 0.6 mgC/(g OC·d) at the beginning to 0.2 mgC/(g OC·d) at the end of incubation. They strongly recommended analyzing the whole time series of soil respiration. Leirs *et al*. (1999), reported a study on C mineralization of samples from Ah and OA horizons of acid forest soils by incubation for 10 d. Their C mineralization rate varied widely from 0.08 to 4.5 mgC/(g OC·d), depending on the interaction of moisture and temperature conditions. The mean C mineralization rate here, ranged from 0.13 mg CO_2-C/(g OC·d) to 0.16 mg CO_2-C/(g OC·d) under 20°C and from 0.17 to 0.20 mg CO_2-C/(g OC·d) under 25°C respectively, that is, 30% to 70% higher than the results obtained by Zhou *et al*. (2003), who studied topsoil C mineralization of three different paddy soils from Southern Jiangsu by incubation under moist condition for 112 d. They found the C mineralization rate ranging from 0.03 to 0.13 mgC/(g OC·d) under 20°C, and 0.11 to 0.13 mgC/(g OC·d) under 25°C, varying with the soil types. However, these rates were lower than the above-mentioned rates for the grassland and forest soils. Apparently, SOC of the studied paddy soils seemed more stable than those of the grassland and forest soils. This could be partly supported by the previous findings that the paddy soils of China had a higher SOC density and a higher C sequestration potential and present rates than the dry cropland soils, American and European grasslands, and forest soils (Pan *et al*., 2003, 2004, 2005). The higher C mineralization rates in the present study than those reported by Zhou *et al*. (2003), may attribute to the higher moisture content during the incubation. The effect of moisture on C mineralization has been well documented. Lomander *et al*. (1998), observed a positive effect of water content on the soil CO_2 respiration of grass soil under laboratory incubation. Leirs *et al*. (1999), had shown that C mineralization rate of forest soils was in linear response to soil moisture. Recently, Borken *et al*. (2003), reported that the soil CO_2 evolution of O horizon of a forest soil was in strong correlation with the soil water content in the field. As the paddy soils are usually in water saturated or wet conditions, higher C mineralization rates could be expected in wet field conditions than under moist laboratory conditions.

The Q_{10} values for forest soils ranged from 2.5 to 4.0, as reported by Leirós *et al*. (1999), and 1.8 to 2.5 of widely varied soil types of forest soils by Smith (2003). They inferred that the temperate forest soils were very sensitive to soil warming with regard to CO_2 evolution. In the present study, however, the Q_{10} values of C mineralization for a total 114 d and during the period from 75 d to 114 d ranged from 1.0 to 2.3 and 1.0 to 2.4 respectively. The Q_{10} values and the C mineralization rates of these paddy soils seemed to be lower than the data for forest and grassland soils mentioned above. And the values of Q_{10} were also generally lower than those reported for uncultivated and cultivated wetland soils from the Sanjing Plain, Heilongjiang, China (Zhang *et al*., 2005). Therefore, the fact that the paddy soils had a higher C density than the dry upland soils and a significant C sequestration trend was detectable in most of the China paddy soils (Pan *et al*., 2003, 2004, 2005) may be attributed more to the relatively great stability of the accumulated SOC . Although both the C mineralization rate and the temperature dependence on C mineralization of the paddy soils varied with the soil types of different origins, the Q_{10} value was a negatively exponential function of the soil initial C/N (Fig.3). Enrichment of soil N could increase the soil microbial respiration in forest soils, possibly because of the formation of the soil microbial community that could actively respond to the added C (Joergensen and Scheu, 1999). Henriksen and Breland (1999), reported that the concentration of available N significantly affected the added OM mineralization with the changes of microbial community.

The N enrichment has been heavily reported in China paddy soils (Xu, 2001; Li *et al*., 2003). The result here also implicates that the N enrichment in paddy soils will potentially raise the sensitivity of C mineralization to global warming. However, the C mineralization within the first 23 d was not found to be correlated to the C/N ratio, indicating that the high availability of C responded to warming rather than the microbial metabolic activity. In addition, a significant correlation of the Q_{10} quotients with the ratio of MBC to TOC was observed (Fig.4). Thornley

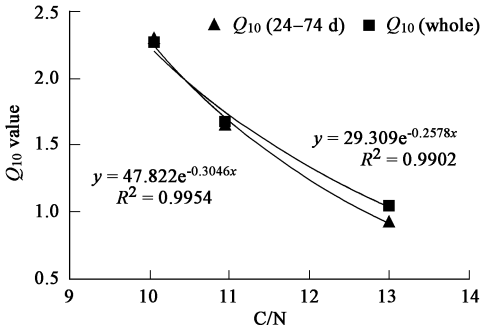

Fig. 3 Correlation of Q_{10} value of C mineralization of the total and during the 24–74 d period of incubation with soil C/N ratio.

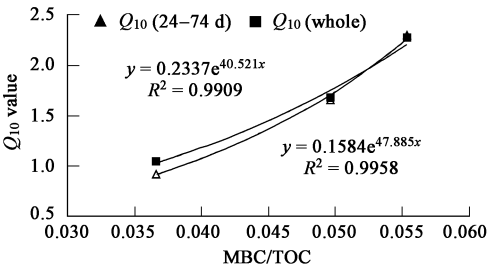

Fig. 4 Correlation of Q_{10} quotient with the MBC/TOC ratio of the soils studied.

et al.(2001), had already demonstrated that the microbial activity might have a more prompt response to warming than the availability of the C substrate. The higher Q_{10} and temperature dependence of the soil EP may be attributed to the higher sensitivity of the microbial activity to warming. However, the variation of the microbial community and the role of C mineralization in these soils are still not clear.

3.2 C pools and the C mineralization of the paddy soils

Many studies on C mineralization from the field and laboratory incubation dealt with first-order decay models, especially with short periods of soil incubations (Collins *et al*., 2000; Henriksen and Breland, 1999; Leirós *et al*., 1999). More recently, Reichstein *et al*. (2000), argued that the temperature dependence was a nonlinear function of incubation time. It was clear that the C mineralization curve of all the three soils under a single temperature did not follow a first order decay of SOM as generally described (Gregorich and Janzen, 2000), for a first order decay equation could only describe 55% to 84% of the C mineralization dynamics of the last 90 d (regression equation not shown) in this study. The fact that C mineralization varied with incubation days indeed implied different C pools, which differed in C bioavailability in these soils in the course of incubation. Côté *et al*. (2000) attributed the difference in C mineralization rates within the first 100 d of incubation from different forest stands to the difference in the labile C pool. Although the model of three pools of soil C is commonly accepted, Collins *et al*. (2000) has shown that the active pool comprised of 3%–8% of total C, slow pool of 50%, and a resistant pool of another half for topsoil under natural vegetation. The total mineralized C after 114 d incubation was assumed as the total mineralizable carbon pool of these soils. The dynamics of C mineralization suggested three subpools for this mineralizable C: a very active subpool available for about three weeks, followed by an active subpool of considerable size lasting for seven weeks, and a slow pool with low and more or less constant rate of C mineralization in the last four weeks. This suggested that a laboratory incubation study of C mineralization from paddy soils could be undertaken for less than 10 weeks. The proportion of a very actively mineralized pool ranged from 24.5% to 28.8%, that of an actively mineralized pool from 46.9% to 50.0% and that of a slowly mineralized pool from 21.1% to 28.3% of the total mineralized C pool in the total 114 d of incubation respectively, slightly varying with the soils. Although the very active subpool exerted a significantly higher mineralization rate and the active subpool gave a dominant contribution to the total C mineralization and the effect of warming, the proportion of the slowly mineralized pool was slightly increased under warming (Table 5 and also see Fig.3). This implicates a relative accessibility of the slow C pool under warming, which was argued in Knorn's (2005) study recently.

Neither the total nor the labile carbon pool could account for the total mineralizable C pool of these soils (Fig.5). Although the very actively and actively mineralizable C under 25°C seemed dependent on the pool of labile C, this dependence was increasingly insignificant for the C mineralized after 23 d of incubation (Fig.6). Contribution by slow C pool should have been involved in the relatively long and slowly mineralizing stage, which could be supported by the heavier C released after 23 d incubation under a single temperature (Table 6). Townsed *et al*. (1997), had discussed that the long stage of slow and stable mineralization could represent the slow C pool of the soils. POC pool, commonly accepted as a slow C pool, was correlated to the actively mineralized C pool rather than to the slowly mineralized pool.

Thus, the accumulation of this slow pool C in the paddy soils (Zhou *et al*., 2006; Peng *et al*., 2004) does not seem to contribute proportionally to the mineralization and the warming effect. This C pool is generally considered as physically protected in macro-aggregates (Six *et al*., 2002), and is shown as not readily accessible to microbial mineralization even under warming (Garten *et al*., 1999). Many studies have demonstrated that the C sequestration in paddy soils is characterized by the increase of SOC in physically protected coarse aggregates in the size of sand particles (Li *et al*., 2007; Yuan *et al*., 2004).

Hence, C mineralization of paddy soils depends not only on the chemical lability of SOC (pool distribution), but also on the microbial metabolic activity and the soil N status. Although labile C may give significant contribution to the very actively and actively mineralizable C, accumulation of younger or labile C does not necessarily enhance the C mineralization potential, which could be considered as a result of mutual interaction of C availability, accessibility to the protected labile C pools, and the metabolic activity of microbes affected by soil nutrient and moisture regimes.

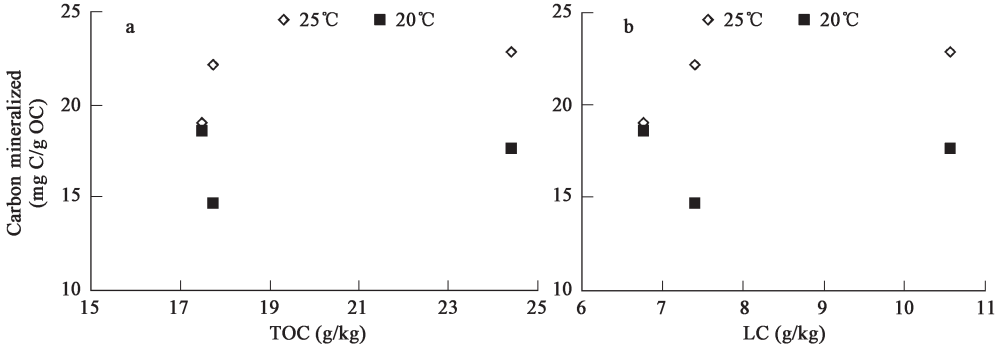

Fig. 5 Correlation of total mineralized C with total (a) and labile (b) carbon of the soils under 20 and 25°C respectively.

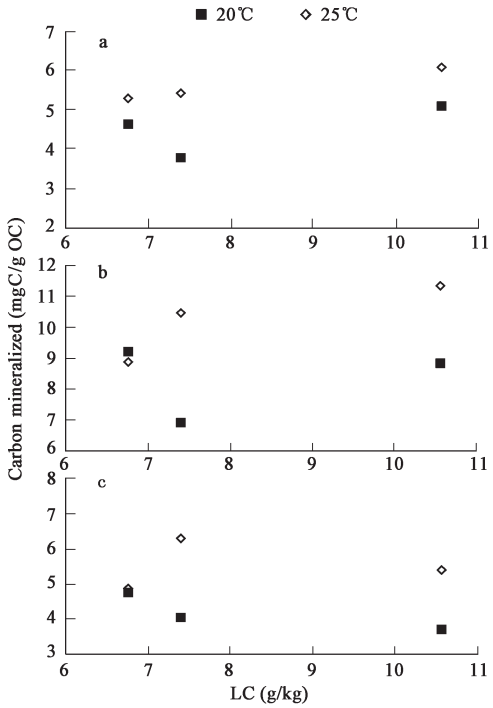

Fig. 6 Correlation of mineralizable C pools with the labile C pool of the studied soils. A very actively mineralized subpool (a); actively mineralized subpool (b), and slowly mineralized subpool (c).

4 Conclusions

The mean C mineralization rate of the paddy soils studied and the temperature dependence was relatively small compared to those of the natural forest soils and grassland soils. This could support the previous finding of the SOC enhancement in the paddy soils of China over the last decades. C mineralization and Q_{10} values of the paddy soils varied with soil types and soil N status. The labile C played an important role in the mineralization of SOC in paddy soils and the warming effect. Moreover, the data here indicated that there were different subpools of the mineralizable C in the paddy soils that had different accessibility to mineralization and different responses to warming. Soil N status also had a controlling role in the C mineralization of the paddy soils although the mineralization rate was relatively lower than the reported values for grassland and dry cropland soils.

The role of C availability and microbial community in the C mineralization of these paddy soils still deserves further study.

Acknowledgements: The authors are grateful to the anonymous reviewers for their suggestions in improving both the scientific and linguistic quality of the manuscript.

References

Blair G J, Lefroy R D B, Lisle L, 1995. Soil carbon fractions based on their degree of oxidation and the development of a carbon management index[J]. Aust J Agric Res, 46: 1459–1466.

Borken W, Davidson E A, Savage K *et al*., 2003. Drying and wetting effects on carbon dioxide release from organic horizons[J]. Soil Sci Soc Am J, 67: 1888–1896.

Bowden R D, Newkirk K M, Rullo G M, 1998. Carbon dioxide and methane fluxes by forest soil under laboratory controlled moisture and temperature conditions[J]. Soil Biology and Biochemistry, 30: 1591–1597.

Cambardella C A, Elliott E T, 1992. Particulate soil organic-matter changes across a grassland cultivation sequence[J]. Soil Science Society of America Journal, 56: 777–783.

Collins H P, Elliott E T, Paustian K *et al*., 2000. Soil carbon pools and fluxes in long-term corn belt agroecosystems[J]. Soil Biology and Biochemistry, 32: 157–168.

Coleman D C, Anderson R V, Cole C V *et al*., 1978. Tropic interactions in soils as they affect energy and nutrient dynamics. IV. Flows of metabolic and biomass carbon[J]. Microbial Ecol, 4: 373–380.

Côté L, Brown S, Paré D *et al*., 2000. Dynamics of carbon and nitrogen mineralization in relation to stand type, stand age and soil texture in the boreal mixedwood[J]. Soil Biology and Biochemistry, 32: 1079–1090.

Gao X, Ding Y, Zhao Z *et al*., 2003. Climate change due to greenhouse effects in China as simulated by a regional climate model part II: climate change[J]. Acta Meterologica Sinica, 61: 29–38.

Garten-Jr C T, Post W M, Hanson P J *et al*., 1999. Forest soil carbon invertories and dynamics along an elevation gradient in the southern appellation mountains[J]. Biogeochemistry, 45: 115–145.

Giardina C P, Ryan M G, 2000. Evidence that decomposition rates of organic carbon in mineral soil do not vary with temperature[J]. Nature, 404: 858–861.

Gong Z, 1999. Chinese soil taxonomic classification: theory, methodology and practices[M]. Beijing: China Science Press. 109–194.

Goulden M L, Wofsy S C, Harden J W *et al*., 1998. Sensitivity of boreal forest carbon balance to soil thaw[J]. Science, 279: 214–217.

Gregorich E G, Janzen H H, 2000. Microbially mediated processes: decomposition[M]. In: Handbook of soil science (Sumner M. F., ed.). New York, Washington D.C., London: Boca Raton. C107–C119.

Henriksen T M, Breland T A, 1999. Nitrogen availability effects on carbon mineralization, fungal and bacterial growth, and enzyme activities during decomposition of wheat straw in soil[J]. Soil Biology and Biochemistry, 31: 1121–1134.

Houghton J T, Meira Filho L G, Bruce J *et al*., 1995. Climate change 1994: Radiative forcing of climate change and an evaluation of the IPCC IS92 emission scenarios. Reports of working groups I and III of the intergovernmental panel on climate change[M]. Cambridge: Cambridge University Press.

Houghton J T, Meira Filho L G, Callendar B A *et al*., 1996. Climate change 1995: The science of climate change[M]. Contribution of working group I to the second assessment report of the intergovernmental panel on climate change. Cambridge: Cambridge University Press.

IPCC, 2001. Climate change 2001: The scientific basis[EB]. Available online at http://www.grida.no/climate/ipcc_tar/wgl/index.htm (verified 18 Nov. 2003).

Jenkinson D S, Adams D E, Wild A, 1991. Model estimates of CO_2 emissions from soil in response to global warming[J]. Nature, 351: 304–306.

Jenkinson D S, Powlson D S, 1976. The effects of biocidal treatments on metabolism in soil. V. A method for measuring soil biomass[J]. Soil Biology Biochemistry, 8: 209–213.

Joergensen R G, Mueller T, Wolters V, 1996. Total carbohydrates of the soil microbial biomass in 0.5 M K_2SO_4 soil extracts[J]. Soil Biology and Biochemistry, 28: 1147–1153.

Joergensen R G, Scheu S, 1999. Response of soil microorganisms to the addition of carbon, nitrogen and phosphorus in a forest Rendzina[J]. Soil Biology and Biochemistry, 31: 859–866.

Kirschbaum M U F, 1995. The temperature dependence of soil organic matter decomposition, and the effect of global warming on soil organic C storage[J]. Soil Biology and Biochemistry, 27: 753–760.

Kirschbaum M U F, 2000. Will changes in soil organic carbon act as a positive or negative feedback on global warming?[J]. Biogeochemistry, 48: 21–51.

Knorr W, Prentice I C, House J I *et al*., 2005. Long-term sensitivity of soil carbon turnover to warming[J]. Nature, 433: 298–300.

Kruse J S, Kissel D E, Cabrera M L, 2004. Effects of drying and rewetting on carbon and nitrogen mineralization in soils and incorporated residues[J]. Nutrient Cycling in Agroecosystems, 69: 247–256.

Leirós M C, Trasar-Cepeda C, Seoane S *et al*., 1999. Dependence of mineralization of soil organic matter on temperature and moisture[J]. Soil Biology and Biochemistry, 31: 327–335.

Li L, Pan G, Zhang X *et al*., 2000. Stable isotopic composition of organic carbon in soil aggregates[J]. Journal of Nanjing Agricultural University, 23: 114–116.

Li Q, 1992. Paddy soils of China[M]. Beijing: Chinese Science Press. 95–101.

Li Z, Li D, Zhang T *et al*., 2003. Dynamics of soil properties of paddy fields in red soil region[J]. Acta Pedologica Sinica, 40: 870–878.

Li Z, Pan G, Zhang X, 2007. Changes in pool distribution and ^{13}C natural abundance of organic carbon of a paddy soil after corn cultivation for 3 years[J]. Acta Pedologica Sinica (in press).

Lomander A, Kätterer T, Andrén O, 1998. Carbon dioxide evolution from top- and subsoil as affected by moisture and constant and fluctuating temperature[J]. Soil Biology and Biochemistry, 30: 2017–2022.

Lu R, 2000. Methods of soil and agrochemical analysis[M]. Beijing: China Agricultural Science and Technology Press.

MacDonald N W, Randlett D L, Donald Z R, 1999. Soil warming and carbon loss from a lake states spodosol[J]. Soil Science Society of America Journal, 63: 211–218.

Mehra O P, Jackson M L, 1960. Iron oxide removed from soils and clays by a Dithionite-citrate system buffered with sodium bicarbonate[J]. Clays and Clay Mineral, 7: 317–329.

Oechel W C, Hastings S J, Jenkins M *et al*., 1993. Recent change of Arctic tundra ecosystems from a net carbon sink to a source[J]. Nature, 361: 520–523.

Oechel W C, Vourlitis G L, Hastings S J *et al*., 1995. Change in Arctic CO_2 flux over two decades: Effects of climate change at Barrow, Alaska[J]. Ecological Applications, 5: 846–855.

Pan G, Li L, Zhang X *et al*., 2003. Soil organic carbon storage of China and the sequestration dynamics in agricultural lands[J]. Advance in Earth Science, 18: 609–618.

Pan G, Li L, Wu L *et al*., 2004. Storage and sequestration potential of topsoil organic carbon in China's paddy soils[J]. Global Change Biology, 10: 79–92.

Pan G, Li L, Zhang Q *et al*., 2005. Organic carbon stock in topsoil of Jiangsu Province, China, and the recent trend of carbon sequestration[J]. J Environ Sci, 17: 1–7.

Peng X, Zhang B, Zhao Q, 2004. A review on relationship between soil organic carbon pools and soil structure stability[J]. Acta Pedologica Sinica, 41: 618–623.

Reichstein M, Bednorz F, Broll G *et al*., 2000. Temperature dependence of carbon mineralization: conclusions from a long-term incubation of subalpine soil samples[J]. Soil Biology and Biochemistry, 32: 947–958.

Schimel J P, Gulledge J M, Clein-Curley J S *et al*., 1999. Moisture effects on microbial activity and community structure in decomposing birch litter in the Alaskan taiga[J]. Soil Biology and Biochemistry, 31: 831–838.

Six J, Conant R T, Paul E A *et al*., 2002. Stabilization mechanisms of soil organic matter: Implications for C-saturation of soils[J]. Plant and Soil, 241: 155–157.

Smith V R, 2003. Soil respiration and its determinants on a sub-Antarctic island[J]. Soil Biology and Biochemistry, 35: 77–91.

Soil Survey Staff, USDA, 1999. Soil taxonomy: A basic system of soil classification for making and interpreting soil surveys[M]. United States Department of Agriculture. Natural Resources Conservation Service.

SSSSC (State Soil Survey Service of China), 1998. China soils[M]. Beijing: China Agricultural Press.

Valentini R, Matteucci G, Dolman A J *et al*., 2000. Respiration as the main determinant of carbon balance in European forests[J]. Nature, 404: 862–864.

Voroney R P, Winter J P, Beyaret R P, 1993. Soil microbial biomass C and N[M]. In: Soil sampling and methods of analysis. Canadian Society of Soil Science (Carter M. R., ed.). Chelsea: Lewis. 277–286.

Xu Q, 2001. Evolution of soil fertility in relation to its quality in paddy field of the Taihu Lake area[J]. Resources and Environment in the Yangtze Basin. 10: 323–328.

Yuan Y, Li H, Huang Q *et al*., 2004. Effects of different fertilization on soil organic carbon distribution and storage in micro–aggregates of red paddy topsoil[J]. Acta Ecologica Sinica, 24: 2961–2966.

Zhang J, Song C, Yang W, 2005. Temperature sensitivity of soil respiration and its effecting factors in the different land use[J] . Acta Scientiae Circumstantiae, 25: 1537–1542.

Zhou P, Zhang X, Pan G, 2006. Effect of long-term different fertilization on total and particulate organic carbon of a paddy soil an example of Huangnitu from the Tai Lake region, China[J]. Plant Nutrition and Fertilizing Science, 12: 765–771.

Zhou Y, Pan G, Li L *et al*., 2003. Change of organic carbon pools and the responses to soil warming during laboratory incubations under different temperatures of 3 kinds of paddy soils in Tai lake region[J]. Environmental Science, 24: 46–51.