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Larger Q_{10} of carbon decomposition in finer soil particles does not bring long-lasting dependence of Q_{10} on soil texture

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Summary

Soil particle-size fractionation is a reliable approach for the separation of carbon (C) pools with different stabilities. Our previous study found that C decomposition in fine soil particles had greater temperature sensitivity (Q_{10}) than in coarse particles in grassland and forest soils. However, it is not known whether this phenomenon occurs in cropland soil and whether it generally suggests a dependence of Q_{10} on soil texture. We carried out a 107-day incubation of isolated soil particles from cropland soil, including paddy and upland, with contrasting fertilizer applications. The incubation was carried out over three short-term cycles of sequentially changing temperatures between 5 and 30°C at 5°C intervals. The results indicated that C decomposition was faster in the sand (>50 μm) fraction than in the silt (2–50 μm) and clay (<2 μm) fractions. However, Q_{10} was generally larger in the clay fractions than in the other two fractions for all types of cropland soil, which is in accordance with our previous study. This suggested that the passive C pool is more sensitive to climate warming than the labile C pool. Considering the aforementioned Q_{10} pattern across soil particles, we hypothesized that fine-textured soil should have a larger Q_{10} than coarse-textured soil. However, we observed this outcome in the first and second temperature cycles only, but not in the third. In conclusion, a larger Q_{10} in finer soil particles is probably a widespread phenomenon, but it does not bring a long-lasting dependence of Q_{10} on soil texture.

Highlights

- Does cropland C flux in finer soil particles have larger Q_{10} as in grassland and forest?
- Larger Q_{10} in clay fractions than in sand particles is widespread.
- The passive C pool is more sensitive to climate warming than the labile C pool.
- Larger Q_{10} in finer soil particles does not bring long-lasting dependence of Q_{10} on soil texture.

Introduction

Soil is the largest global terrestrial carbon (C) reservoir, which is more than two or three times the size of the vegetation or atmospheric C reservoirs. The decomposition of soil organic C is the largest carbon flux from terrestrial systems to the atmosphere (Schlesinger & Andrews, 2000). Climate warming will accelerate soil C decomposition and cause soil to emit a larger amount of CO_2 into the atmosphere (Bond-Lamberty & Thomson, 2010). This may further increase the Earth's temperature; thus, soil C decomposition could act as positive feedback to climate warming (Davidson & Janssens, 2006). The temperature sensitivity of soil C decomposition, usually termed Q_{10} (the proportionate increase in the reaction rate for a warming of 10 K), is an important parameter in current models for the interaction of the global C cycle and climate change.

Soil contains differently sized mineral particles, which are usually classified simply as sand, silt and clay. Clay and silt particles contain mainly sesquioxides and layer silicates, and provide large specific surface areas and numerous reactive sites (hereafter called sorption capacity) at which SOM can be sorbed by strong ligand exchange and polyvalent cation bridges. Sand particles are dominated by

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quartz and have a weak bonding affinity only (hereafter called sorption strength) to SOM (Christensen, 2001; von Lützow *et al*., 2007). Empirical evidence has shown that soil C in the clay and silt fractions is more stable than that in the sand fraction (e.g. Bimüller *et al*., 2014; Ding *et al*., 2014). Therefore, C within the sand fraction is allocated to the active pool, and C in silt and clay fractions is allocated to the intermediate and passive pools (von Lützow *et al*., 2007). Particle-size separation with subsequent incubation experiments provides a useful approach to describe soil C stability within discrete fractions obtained from a complex matrix (Bimüller *et al*., 2014).

Apart from the difference in C stabilities, these soil particles might also vary in Q_{10}. There are two reasons for this speculation. First, these particles have different substrate availability (Nelson *et al*., 1994) (i.e. the substrate concentration at the active site of the enzyme, which is a key regulatory factor in Q_{10}) (Davidson & Janssens, 2006). Second, SOM molecules in these particles have different degrees of recalcitrance (Christensen, 2001). Here, C recalcitrance is specifically defined as the molecular-level characteristics of organic substances (not including soil matrix protection or climate limitation), which affect their degradation by microbes and enzymes (Sollins *et al*., 1996). According to kinetic theory, more recalcitrant SOM should be more sensitive to temperature (Bosatta & Ågren, 1999).

Investigation of Q_{10} in soil particles can reveal whether the passive or labile C pool is more sensitive to global warming. An answer to this question is a prerequisite for understanding the long-term response of global soil C stocks under climate warming (Fang *et al*., 2005). Furthermore, investigation of Q_{10} in discrete soil particles might disclose the relation between Q_{10} and soil texture. If C in finer soil particles has a larger Q_{10}, C cycling in finer-textured soils might be more sensitive to climate warming, and *vice versa*. Although several studies have explored the relation between Q_{10} and soil texture (Hamdi *et al*., 2013; Ding *et al*., 2016), the outcome is elusive.

There has been no previous research on Q_{10} for differently-sized particle fractions, except for our previous study (Ding *et al*., 2014). We found that C decomposition had a larger Q_{10} in fine than in coarse soil particles in grassland and forest soils (Ding *et al*., 2014). However, we do not know whether this phenomenon occurs also in croplands with different management systems. For instance, paddy and upland cropping areas have different water managements, resulting in contrasting soil microenvironments (moisture, aeration conditions, and so on) and C sequestration mechanisms (Ouyang *et al*., 2014). Furthermore, cropland soils at different places receive different types of fertilizer treatments (e.g. chemical and organic fertilizer), which can alter the amount and molecular composition of C within soil particle-size fractions (Schulten & Leinweber, 1991).

We carried out a 107-day incubation of sand (>50 μm), silt (2–50 μm) and clay (<2 μm) fractions from one paddy soil and one upland soil with organic or inorganic fertilizer treatments. The soil samples were subjected to three short-term cycles including sequential increases and decreases in temperature between 5 and 30°C at an interval of 5°C. The Q_{10} value was estimated from the dynamics of the rates of C decomposition under varying temperatures. We explored the relation between the Q_{10} of bulk soil and clay and sand contents with data from this study and our previous study for grassland and forest bulk soils (Ding *et al*., 2014). Our objectives were to test two hypotheses: (i) soil C decomposition in fine soil particles is more sensitive to warming than in coarse particles of cropland soil, regardless of water and fertilizer managements, and (ii) C decomposition in fine-textured soil will be more sensitive than in coarse-textured soil.

Materials and methods

Sites and sampling

Soil samples were collected in 2013 from two cropped fields, including one upland (no regular submergence throughout the year, except for temporary flooding during irrigation or after heavy rain events) and one paddy field (submerged during the year for rice cultivation). The upland field was at the eco-farming research station of Shandong Agricultural University (35°26′N, 117°49′E, 350 m a.s.l.) in Pingyi County, Shandong Province, China. The mean annual temperature (MAT) is 13.5°C, and the mean annual precipitation (MAP) is 772 mm, with 211 frost-free days. The crops were winter wheat (*Triticum aestivum* L.) and summer maize (*Zea mays* L.) in a yearly rotation system. The soil belongs to the Alfisols order according to the USDA Soil Taxonomy (Soil Survey Staff, 1999). The fertilizer experiment began in 2009. We sampled soil where inorganic fertilizer (NPK) only and organic manure (cow manure, OM) only had been applied; they provided an equal amount of total nitrogen input (375 kg hm^{-2} $years^{-1}$). The soil under NPK treatment was composed of 43.2% sand, 38.8% silt and 18.0% clay, with a pH of 5.3, and that under OM treatment comprised 42.9% sand, 38.0% silt and 19.1% clay, with a pH of 6.3. The paddy field was at a long-term monitoring station of the Ministry of Agriculture (28°16′N, 112°33′E, 48 m a.s.l.) in Ningxiang County, Hunan Province, China. The MAT and MAP are 17.2°C and 1553 mm, respectively, with 247 frost-free days. The rotation system is early and late rice (*Oryza sativa* L.) and winter barley (*Hordeum vulgare* L.) during 1 year. The soil belongs to the Entisols order (Soil Survey Staff, 1999). The fertilizer experiment began in 1986. We sampled the soil to which NPK only and organic manure (pig manure) plus NPK (MNPK) were applied; the two treatments have an equal amount of total nitrogen input (530 kg hm^{-2} $years^{-1}$). In the MNPK treatment, 40% of the nitrogen was from inorganic fertilizer and the other 60% was from organic manure. The soils were composed of 17.3% sand, 49.7% silt and 33.0% clay, with a pH of 6.5, and 17.5% sand, 54.0% silt and 28.5% clay, with a pH of 6.6, for the NPK and MNPK treatments, respectively. The soil texture and pH were measured by the hydrometer method and the electrode method, respectively.

For each treatment, 40 cores of mineral soil at 0–20-cm depths were collected randomly using an auger of 4-cm diameter and mixed to form one sample. A small fraction of the soil was placed immediately in a sterile plastic bag on ice for transport to the laboratory and stored at 4°C for making inoculum for incubation.

The remaining soil samples were air-dried at room temperature and then crumbled and sieved through a 2-mm mesh to remove visibly identifiable plant debris and gravel before particle-size fractionation.

Particle-size fractionation

The bulk soil samples were divided into clay, silt and sand fractions through a combination of sonic dispersion, wet sieving and centrifugation (Tiessen & Stewart, 1983; Ding *et al*., 2014). This procedure is different from soil texture measurement (hydrometer method) that can determine soil particle-size distribution only, but does not separate these particles. Specifically, every 20 g of dried soil was dispersed in 100 ml of deionized water with an ultrasonic probe (6-mm diameter and 300 W), JY92-II Ultrasonic Homogenizer (Ningbo Scientz Biotech, Ningbo, China). The ultrasonic time was chosen as 2 and 15 minutes for the upland and the paddy soils, respectively, after calibration against their soil textures in our preliminary experiment because different soils require different ultrasonic energy inputs (Ding *et al*., 2014). The sand fraction (>50 μm) was separated by washing all of the fine particles in the soil suspension through a 50-mm sieve until the elutriate solution was clear. The elutriate solution was divided into silt (2–50 μm) and clay (<2 μm) fractions by repeated centrifugation with an Avanti J-25 centrifuge (Beckman Coulter, Brea, CA, USA). The isolated soil fractions were oven-dried at 40°C and were crushed gently for incubation. The organic C concentrations were determined by the $K_2Cr_2O_7$–H_2SO_4 (potassium dichromate–sulphuric acid) wet oxidation method. The total nitrogen (TN) and phosphorus (TP) concentrations were measured by the Kjeldahl method and ammonium molybdate spectrophotometric method after sodium hydroxide fusion, respectively.

Laboratory incubation

The soil particle-size fractions (clay, silt and sand) and the bulk soil, each comprising a 10-g dried sample, were placed in 150-ml brown flasks with 12 replicates. Then, 2 ml of inoculum was added to each sample. The inoculum was made by shaking 100 g of corresponding fresh bulk soil (kept at 4°C) with 1000 ml of deionized water; this was left to stand overnight for it to clear. Empty flasks were used as blanks. The soil moisture for all samples was adjusted to 60% water-filled pore space (WFPS), at which point microbes are considered to have maximum aerobic activity. The total incubation lasted for 107 days. Three temperature-change cycles were started on days 9, 47 and 100 after the beginning of the incubation, and each cycle lasted for 8 days (Figure S1, Supporting Information). The first temperature-change cycle began after 8 days of pre-incubation to avoid pulses in microbial activity induced by the disturbance of mixing and adjustment of moisture content. The soil samples were kept at room temperature (ca. 20°C) during the period between cycles. Soil moisture was corrected regularly for water loss by adding deionized water every 3 days. Within each temperature-change cycle, four replicate flasks at room temperature were chosen randomly and placed in an incubator with a temperature of 5°C. The temperature was increased continuously from 5 to 30°C in intervals of 5°C, and then gradually decreased from 30 to 5°C at the same intervals. After each temperature change, an equilibration period of 3 hours was required for the incubation flasks to allow the soil samples to adapt to the change in temperature and to avoid possible hysteresis of CO_2 release in response to temperature changes (Chen *et al*., 2010). During the equilibration period, the flasks were covered with porous film to maintain the water content of the soil. Then, the flasks were sealed using butyl rubber septa and flushed with ambient air for 1 minute (enough for entire replacement of the gas in the flask) using a pump with an 'inlet–outlet' needle inserted through the septa to produce the same initial gas conditions among all flasks (including the blanks). After pre-set time lengths of sealed incubation (respiration time: 20, 15, 9, 6, 4 and 3 hours for 5, 10, 15, 20, 25 and 30°C, respectively), 5 ml of gas was extracted from the headspace of each flask through gas-tight injection syringes; then, the rubber septa were opened before the next temperature level. The respiration time was determined in our preliminary experiment to ensure that the CO_2 concentration in the sample flask was considerably larger than its initial CO_2 concentration, but below 1% to avoid influencing the process of soil CO_2 diffusion.

The CO_2 concentration in the syringes was analysed with a gas chromatograph (Agilent 7890A, Santa Clara, CA, USA). The rate of C decomposition was then calculated from the net rate of CO_2 accumulation in the headspace, assuming that CO_2 concentrations increased linearly with incubation time before it exceeded 1%. After each cycle, the four replicates were destructively sampled, and each soil sample was divided into two equal parts, one with fumigation and one without fumigation, to measure microbial biomass carbon (MBC) by chloroform fumigation-extraction (K_2SO_4) with a Multi N/C 3100-TOC/TN Analyzer (Analytik Jena, Jena, Germany). The K_2SO_4-extractable organic C and nitrogen from the non-fumigated soil samples were considered to be the dissolved organic carbon (DOC) and total dissolved nitrogen (TDN).

Calculations and statistical analysis

According to Ding *et al*. (2014), the rate of C decomposition was calculated with the following equation:

$$R = V \times [273/(273+T)] \times (C_s - C_b) \times \rho \times (12/44) / (W \times c \times t), \quad (1)$$

where R is the rate of C decomposition (μg CO_2-C g^{-1} C $hour^{-1}$), V is the flask headspace volume (litre), T is the incubation temperature (°C), C_s and C_b are the CO_2 concentrations (μl l^{-1}) in the sample and blank flasks (representing the initial CO_2 conditions in the sample flask before enclosing incubation), respectively, ρ is the standard state CO_2 density (44/22.4 = 1.96 g l^{-1}), W is the dry weight of the soil particles (kg), c is the C concentration of the soil particles (g kg^{-1}) and t is the time enclosed (hour). The final rate of C decomposition was calculated as the average of the values at a given temperature between the increasing and decreasing temperature

periods. We used a first-order exponential model to fit the relation between R and temperature:

$$R = A \cdot e^{k \cdot T}, \quad (2)$$

where and A and k are the model parameters that are to be determined. The Q_{10} values were then computed as follows:

$$Q_{10} = e^{10 \times k}. \quad (3)$$

We used a one-way analysis of variance (ANOVA) to test the difference in the DOC, TDN, MBC, R at 20°C and the Q_{10} among the particle-size fractions for each soil type for a given temperature change cycle. Where significant differences ($P < 0.05$) were observed, the least significant difference (LSD) at 5% probability was used to assess the significance of differences between soil particle-size fractions. We used a two-way ANOVA with factors including temperature change cycles and soil particles (clay, silt and sand) to test the effect of incubation time and its interaction with particle-size fractions on the soil variables of a given type of soil. Normality of residuals and homogeneity of the variances of the residuals across groups were checked for each ANOVA. Relations between the above soil variables were determined by correlation analyses. The relations between Q_{10} and clay and sand contents were analysed with the combined data from this and our previous study (Ding *et al*., 2014). We chose the statistical significance to be $P = 0.05$.

To gain a mechanistic understanding of the difference in Q_{10} among soil particle-size fractions, we used structural equation modelling to explore the direct and indirect effects of the diameter of soil particles on C decomposition and Q_{10}. Structural equation modelling is a multivariate statistical analysis technique based on a collection of simultaneous procedures that test the hypothetical pathways of influence (direct and indirect) among many variables using the covariance among those variables (Grace, 2006). This method relates predictors directly to the response variable and disregards the overall effects derived from the interactions among variables, thereby going beyond traditional multivariate techniques (Grace, 2006).

To quantify the diameter of the soil particles, we calculated the index of geometric particle-size diameter (PSD, d_g) according to Shirazi & Boersma (1984). The PSD of bulk soil is calculated as follows:

$$d_g = \exp\left(0.01 \times \sum_{i=1}^{n} f_i \left(\ln M_i\right)\right), \quad (4)$$

where n is the number of separate soil groups (e.g. clay, silt and sand), f_i is the percent of total soil mass with a diameter equal to or less than M_i and M_i is the arithmetic mean of two consecutive particle-size limits. Thus, for the USDA limits, M_{clay} is $(0 + 0.002)/2 = 0.001$ mm, M_{silt} is $(0.002 + 0.05)/2 = 0.026$ mm and M_{sand} is $(0.05 + 2.00)/2 = 1.025$ mm.

We carried out the structural equation modelling separately for the influencing mechanisms in R and Q_{10} based on data on bulk soil and soil particles during the first and third cycles because we did not measure soil DOC and TDN contents in the second cycle. For the structural equation modelling we specified a conceptual model of hypothetical relations based on *a priori* theoretical knowledge. We assumed that soil particle size had a direct effect on R and Q_{10}. We also assumed that differently sized soil particles had different degrees of substrate C and N availability and C recalcitrance, which in turn would affect R, and different degrees of C availability and recalcitrance would affect Q_{10}. Soil C availability was indicated by the concentration of DOC, according to Wang *et al*. (2003). Soil N availability was represented by the concentration of TDN. Carbon recalcitrance was represented by the C/N ratio, which is a good indicator of the decomposability of plant litter (Tian *et al*., 1992). Although there is no direct indicator for the capacity and strength of sorption of SOM on mineral surfaces, we assumed that they played a critical role in the direct effect of soil particle size on soil properties because sorption status is closely linked with size of particle (Christensen, 2001). Next, we examined the modification indices to ensure that no important paths were left out of the model. The adequacy of the models was determined with a chi-squared (χ^2) test, the Akaike information criterion (AIC) and root square mean errors of approximation (RMSEA) (Grace, 2006). Adequate model fits are indicated by a non-significant χ^2 test ($P > 0.05$), small AIC and small RMSEA (<0.05), which suggest that there is a small difference only between the modelled and observed values (Grace, 2006). The coefficient of each causal relation was expressed as a standardized path coefficient, which is based on variance relations and is the amount of change in standard deviations of the dependent variable per standard deviation change in the independent variable. The standardized path coefficient can represent the magnitude of the direct effect of the predictor on the outcome. The magnitude of the indirect effect of a predictor was obtained by multiplying all of the path coefficients connecting the predictor to our response variable (Grace, 2006). Structural equation modelling was carried out with AMOS 21.0 software (IBM, SPSS, Armonk, NY, USA). The other statistical analyses were performed with the SPSS 17.0 package (SPSS, Chicago, IL, USA).

Results

Soil chemical characteristics

The upland and paddy soils had different patterns of C, N and P stoichiometry across the soil particle-size fractions (Table 1). In the upland soil, the C, TN and TP concentrations decreased with increasing particle size (maximum in the clay fraction and minimum in the sand fraction). However, in the paddy soil, the concentrations of these elements were at a minimum in the medium-sized fraction (silt fraction). The C/N ratios increased with increasing particle size in the paddy soil but were smallest in the medium-sized fraction in the upland soil. The OM or MNPK treatment had greater C, TN and TP concentrations than the NPK treatment for all of the soil fractions and bulk soil samples.

There were significant differences ($P < 0.001$) in the DOC and TDN contents among the different soil fractions within each soil

Table 1 Organic carbon (C), total nitrogen (TN) and phosphorus (TP) status in different particle-size fractions of cropland soil

		Soil particle-size fractions and bulk soil			
Characteristics	Fertilizer	Clay (<2 μm)	Silt (2–50 μm)	Sand (>50 μm)	Bulk soil
Upland soil					
Fractionation[a] / %	NPK	16.2	37.9	43.3	97.4
	OM	16.2	37.9	43.3	97.4
C/g kg^{-1} / %[b]	NPK	17.3 (38.3)	6.8 (35.3)	4.7 (27.7)	7.3
	OM	21.4 (32.5)	7.3 (25.9)	10.5 (42.7)	10.6
TN/g kg^{-1} / %[b]	NPK	1.4 (30.5)	0.7 (35.8)	0.3 (17.2)	0.7
	OM	1.8 (25.1)	0.8 (27.7)	0.8 (31.7)	1.2
TP/g kg^{-1} / %[b]	NPK	0.9 (29.1)	0.4 (31.9)	0.4 (33.2)	0.5
	OM	1.1 (29.7)	0.4 (26.4)	0.6 (40.6)	0.6
C/N	NPK	12.4	9.7	15.9	9.9
	OM	11.9	8.6	12.4	9.2
N/P	NPK	1.6	1.7	0.8	1.5
	OM	1.6	2.0	1.5	1.9
Paddy soil					
Fractionation[a] / %	NPK	35.5	46.1	16.3	97.9
	MNPK	26.5	53.6	18.3	98.4
C/g kg^{-1} / %[b]	NPK	25.4 (52.6)	9.3 (25.0)	24.8 (23.6)	17.2
	MNPK	41.9 (33.0)	24.1 (38.3)	44.4 (24.2)	33.6
TN/g kg^{-1} / %[b]	NPK	2.9 (56.2)	0.9 (22.8)	1.1 (9.9)	1.8
	MNPK	4.4 (38.2)	2.0 (34.6)	2.8 (16.6)	3.1
TP/g kg^{-1} / %[b]	NPK	0.8 (51.9)	0.3 (25.2)	0.3 (9.4)	0.6
	MNPK	3.2 (32.2)	1.4 (29.0)	3.4 (23.4)	2.6
C/N	NPK	8.9	10.4	22.6	9.5
	MNPK	9.4	12.1	15.9	10.9
N/P	NPK	3.5	2.9	3.4	3.2
	MNPK	1.4	1.4	0.8	1.2

[a]The values are the recoveries of soil during particle-size fractionation.
[b]The values in brackets are the percentage (%) of C, TN or TP in the soil fractions of the total amount of bulk soil.
NPK, inorganic fertilizer only; OM, organic manure (cow manure); MNPK, organic manure (pig manure) plus NPK.

type (Table 2). The patterns of DOC and TDN contents across the soil particles were almost the same as those of C and TN contents. They decreased with increasing particle size for the upland soil, but were at a minimum in the silt of the paddy soil. The MBC was significantly different ($P < 0.05$) among the soil particles in the upland and paddy soil with the NPK treatment, but this was not so for the paddy soil with MNPK. The MBC contents of the upland soil were larger in the clay fraction than in the silt and sand fractions.

The contents of DOC, TDN and MBC were larger in the paddy soil samples than in the upland soil, and were generally larger with the manure treatment than with the NPK treatment (Table 2). The incubation time (cycle) had significant effects ($P < 0.01$) on the DOC and TDN contents, but did not have a significant effect ($P > 0.05$) on the MBC content (Table 3). The DOC contents in the majority of soil samples (except for some under the NPK treatment) decreased from the first to third cycle, with the largest decline in the clay fraction (Table 2). Conversely, the TDN contents in all the soil samples increased from the early cycle to the late cycle. The MBC in the majority of soil samples, however, did not change with incubation time, except for the upland bulk soil that showed a decreasing trend.

Rates of carbon decomposition

The rate of soil C decomposition at 20°C (R) was expressed as the CO_2-C rate of emission per unit of soil C (Table 4). There were significant differences in R among the soil particles for all soil types ($P < 0.05$). The sand fraction showed the largest R as expected. Rates of C decomposition at 20°C in the sand fractions averaged 1.8 and 2.1 times those in the silt and clay fractions, respectively. Bulk soil generally had smaller R than the sand fractions, but larger than the clay or silt fractions.

The rates of C decomposition in all of the soil samples were faster in the upland than in the paddy soil for the same temperature-change cycle (Table 4). In the upland soil, most of the soil samples (except for sand) had larger R for the manure treatment than the NPK treatment during a given temperature-change cycle. By contrast, in the paddy soil most of the soil samples had smaller R for the manure treatment. Furthermore, incubation time had a significant

Table 2 Dissolved organic carbon (DOC), total dissolved nitrogen (TDN) and microbial biomass carbon (MBC) in different particle-size fractions of cropland soil

	Cycle 1						Cycle 3					
Soil fractions	DOC / mg kg^{-1}		TDN / mg kg^{-1}		MBC / mg kg^{-1}		DOC / mg kg^{-1}		TDN / mg kg^{-1}		MBC / mg kg^{-1}	
Upland soil	NPK	OM	NPK	OM	NPK	OM	NPK	OM	NPK	OM	NPK	OM
Bulk soil	73 ± 3.4	85 ± 2.6	38 ± 0.8	67 ± 1.8	110 ± 13.7	127 ± 4.0	75 ± 5.9	71 ± 1.7	95 ± 6.1	134 ± 3.0	45 ± 5.0	103 ± 5.8
Clay	171	195	81	138	91	129	118	127	173	295	114	129
Silt	54	51	24	47	35	53	44	51	58	82	39	47
Sand	40	56	15	23	27	92	40	44	33	63	26	81
LSD (5%)	9.2	36.2	3.4	14.9	16.7	44.7	24.6	19.2	20.8	36.6	19.1	59.3
F-ratio$_{(2, 9)}$	617.1	52.3	1135.8	170.6	44.0	7.4	33.0	58.4	131.7	126.5	63.8	5.0
P value	<0.001	<0.001	<0.001	<0.001	<0.001	0.012	<0.001	<0.001	<0.001	<0.001	<0.001	0.035
SE	2.9	11.3	1.1	4.7	5.2	14.0	7.7	6.0	6.5	11.4	6.0	18.5
Paddy soil	NPK	MNPK	NPK	MNPK	NPK	MNPK	NPK	MNPK	NPK	MNPK	NPK	MNPK
Bulk soil	133 ± 6.4	183 ± 6.9	190 ± 1.9	247 ± 13.9	262 ± 30.4	448 ± 35.0	96 ± 4.2	155 ± 11.6	240 ± 6.2	408 ± 32.5	272 ± 14.0	429 ± 16.3
Clay	233	428	315	428	250	319	183	285	511	769	268	308
Silt	53	101	75	90	185	314	64	93	102	149	141	307
Sand	103	159	79	133	254	315	78	133	111	254	256	328
LSD (5%)	9.6	16.8	7.2	18.8	46.6	–	23.5	22.5	16.9	25.0	–	–
F-ratio$_{(2, 9)}$	967.5	1109.0	3787.3	979.6	6.9	0.019	91.8	205.7	2284.8	1804.2	3.0	0.407
P value	<0.001	<0.001	<0.001	<0.001	0.015	0.982	<0.001	<0.001	<0.001	<0.001	0.101	0.677
SE	3.0	5.2	2.2	5.9	14.6	17.4	6.8	7.0	4.9	7.8	40.5	18.4

The data for cycle 2 were not determined. Data of bulk soil are shown as means ± standard error ($n = 4$). F-ratio, P and SE (standard error) values represent the results from a one-way analysis of variance (ANOVA) (degrees of freedom (d.f.) = 2 for the treatment and d.f. = 9 for the error) for DOC, TDN and MBC among different sizes of soil particles. LSD is least significant difference; it was significant for $P < 0.05$. NPK, inorganic fertilizer only; OM, organic manure (cow manure); MNPK, organic manure (pig manure) plus NPK.

effect on R ($P < 0.01$, Table 3), which decreased continuously with incubation time (Table 4).

Temperature sensitivity of C decomposition

Rates of C decomposition had an exponential relation with incubation temperature for all soil samples (Figures S2 and S3, Supporting Information). The temperature coefficient (Q_{10}) showed a significant difference among soil particles for all soil types, except for NPK treatment in the upland soil during the first and second cycles (Table 5). For the paddy soil, Q_{10} decreased with increasing particle size (maximum in clay fraction and minimum in sand fraction). In the upland soil, Q_{10} values were generally larger in the clay fraction than in the silt and sand fractions (except for the NPK treatment during the first and second cycles) and were generally similar for the latter two fractions (Table 5).

The Q_{10} of bulk soil was larger in the paddy than in the upland soil during the first and second cycles, but it was smaller in the paddy soil during the third cycle (Table 5). The Q_{10} of bulk soil was similar for the manure and NPK treatments for both the paddy and upland soils (Table 5). Furthermore, the incubation time and its interaction with soil particle-size fractions had significant effects on Q_{10} ($P < 0.01$, Table 3). The Q_{10} in most upland soil samples increased continuously with incubation time (except for the silt fraction under the NPK treatment); for the paddy soil, however, Q_{10} values in bulk soil decreased from the first and second cycles to the third, and they did not show a regular pattern in the soil fractions (Table 5).

Effect of soil particle size on C decomposition and Q_{10}

The final structural equation model fitted the effect of soil particle-size diameter on R adequately (Figure 1a, $\chi^2 = 2.7$, d.f. = 2, $P = 0.263$, Table S1, Supporting Information). The model described 43% of the variation in R. Soil particle-size diameter had a direct positive effect on R (standardized path coefficient = 0.45, $P < 0.05$), an indirect positive ($-0.31 \times (-0.78) = 0.24$, $P < 0.05$) effect through N availability and negative effects through soil C availability ($-0.26 \times 0.35 = -0.09$, $P < 0.05$) and C recalcitrance ($-0.36 \times 0.80 = -0.29$, $P < 0.05$). Carbon availability had a positive effect on R, but N availability and C recalcitrance exerted negative effects. The absolute value for the total positive effects of soil particle size (0.69) on R was larger than that for the total negative effects (−0.38).

The final structural equation model also fitted the effect of soil particle-size diameter on Q_{10} adequately (Figure 1b, $\chi^2 = 0.2$, d.f. = 1, $P = 0.655$, Table S2, Supporting Information). The model described 45% of the variation in Q_{10}. Soil particle-size diameter did not have a significant direct effect on Q_{10} ($P > 0.05$), but had significant indirect effects on Q_{10} through C availability ($-0.26 \times 0.57 = -0.15$, $P < 0.05$) and recalcitrance

Table 3 Results of two-way analysis of variance (ANOVA) with dissolved organic carbon (DOC), total dissolved nitrogen (TDN), microbial biomass carbon (MBC), rate of C decomposition at 20°C (R) and Q_{10} as independent variables

			Independent variables																	
			DOC				TDN				MBC				R				Q_{10}			
Cropland type	Fertilizer	Source of variation	d.f.	MS	F	P	d.f.	MS	F	P	d.f.	MS	F	P	d.f.	MS	F	P	d.f.	MS	F	P
Upland soil	NPK	Particle	2	26 805	199	<0.01	2	24 409	282	<0.01	2	13 543	108	<0.01	2	700	1836	<0.01	2	0.03	4	0.02
		Cycle	1	2618	19	<0.01	1	14 087	163	<0.01	1	436	3	0.08	2	667	109	<0.01	2	0.11	19	<0.01
		Cycle•Part	2	1538	11	<0.01	2	2971	34	<0.01	2	336	3	0.10	4	25	4	0.01	4	0.04	7	<0.01
		Residual	18	135			18	87			18	126			27	6			27	0.01		
	OM	Particle	2	32 603	99	<0.01	2	71 466	235	<0.01	2	12 594	12	<0.01	2	563	233	<0.01	2	0.23	42	<0.01
		Cycle	1	4082	12	<0.01	1	35 761	117	<0.01	1	199	0.19	0.67	2	1684	696	<0.01	2	0.56	106	<0.01
		Cycle•Part	2	2660	8	<0.01	2	9426	31	<0.01	2	61	0.06	0.95	4	41	17	<0.01	4	0.04	7	<0.01
		Residual	18	328			18	305			18	1078			27	2			27	0.01		
Paddy soil	NPK	Particle	2	42 694	554	<0.01	2	2 361 423	5811	<0.01	2	23 412	6	<0.01	2	479	381	<0.01	2	1.61	500	<0.01
		Cycle	1	2278	30	<0.01	1	37 289	918	<0.01	1	365	0.10	0.76	2	404	322	<0.01	2	0.03	9	<0.01
		Cycle•Part	2	1638	21	<0.01	2	15 716	387	<0.01	2	2116	0.57	0.58	4	40	32	<0.01	4	0.11	35	<0.01
		Residual	15	77			15	41			18	3706			27	1			27	0.003		
	MNPK	Particle	2	151 936	985	<0.01	2	531 794	2782	<0.01	2	246	0.19	0.83	2	98	297	<0.01	2	1.17	393	<0.01
		Cycle	1	21 110	137	<0.01	1	181 915	952	<0.01	1	14	0.01	0.92	2	130	396	<0.01	2	0.049	17	<0.01
		Cycle•Part	2	10 871	70	<0.01	2	44 106	231	<0.01	2	326	0.26	0.78	4	3	10	<0.01	4	0.12	42	<0.01
		Residual	18	154			18	191			18	1278			27	0.3			27	0.003		

Part, particles; d.f., degree of freedom; MS, mean squares; F, variance ratio; $P = F$-probability. Cycle included the first and third temperature change cycles for DOC, DTN and MBC, and the full three cycles for R and Q_{10}.
NPK, inorganic fertilizer only; OM, organic manure (cow manure); MNPK, organic manure (pig manure) plus NPK.

Table 4 Rates of soil C decomposition (R) at 20°C in the cropland soil particle-size fractions

			R / μg CO_2-C g^{-1}C $hour^{-1}$				ANOVA results			
Cropland	Cycles	Fertilizer	Bulk soil	Clay	Silt	Sand	LSD (5%)	F-ratio$_{(2, 9)}$	P	SE
Upland	1	NPK	38.6 ± 0.46	24.8	19.5	34.4	6.2	15.4	0.001	1.9
		OM	46.9 ± 0.72	33.7	29.0	40.1	3.8	22.6	<0.001	1.2
	2	NPK	18.5 ± 0.25	8.5	9.6	21.0	2.4	83.3	<0.001	0.8
		OM	22.3 ± 0.22	10.7	12.7	19.4	4.1	20.6	<0.001	1.3
	3	NPK	15.8 ± 0.67	6.9	10.1	23.9	1.6	340.0	<0.001	0.5
		OM	19.9 ± 0.39	7.8	12.4	21.5	1.4	269.2	<0.001	0.4
Paddy	1	NPK	19.3 ± 0.45	13.3	14.9	30.7	2.4	158.9	<0.001	0.8
		MNPK	15.9 ± 0.40	11.6	10.5	17.6	1.3	91.8	<0.001	0.4
	2	NPK	10.2 ± 0.15	5.9	9.5	16.9	1.5	142.4	<0.001	0.5
		MNPK	9.4 ± 0.09	6.9	7.1	11.6	0.6	206.7	<0.001	0.2
	3	NPK	9.7 ± 0.19	4.9	8.4	12.8	1.2	112.8	<0.001	0.4
		MNPK	8.7 ± 0.42	5.6	5.6	9.3	0.7	83.9	<0.001	0.2

Data of bulk soil are shown as means ± standard error ($n = 4$). F, P and SE (standard error) values represent the result of a one-way ANOVA (degrees of freedom (d.f.) = 2 for the treatment and d.f. = 9 for the error) for R among different sizes of soil particles. LSD is the least significant difference; it is significant for $P < 0.05$. NPK, inorganic fertilizer only; OM, organic manure (cow manure); MNPK, organic manure (pig manure) plus NPK.

($-0.32 \times 0.80 = -0.26$, $P < 0.05$). Carbon availability had a positive effect on Q_{10} (0.57, $P < 0.05$), but C recalcitrance exerted a negative effect (-0.32, $P < 0.05$).

Relation between Q_{10} and soil texture

The paddy and upland soils are classified as silty clay loam and loam, respectively, and the texture of the former was finer than that of the latter. We can see that the silty clay loam (paddy bulk soil) had significantly larger Q_{10} values than the loam (upland bulk soil) in the first and second temperature change cycles only, but were not larger in the third (Table 5). This dependence on incubation time was also evident when we explored the relation between Q_{10} of bulk soil and clay or sand content (Figure 2). During the first temperature change cycle, Q_{10} showed a significant positive linear relation with clay content ($R^2 = 0.78$, $P = 0.02$) and a significant negative one with sand content ($R^2 = 0.68$, $P = 0.04$). However, the coefficient of determination (R^2) decreased continuously with incubation time, and the P values were greater than 0.05 in the second and third cycles.

Table 5 The Q_{10} values of carbon decomposition in the cropland soil particle-size fractions

			Q_{10}				ANOVA results			
Croplands	Cycles	Fertilizer	Bulk soil	Clay	Silt	Sand	LSD (5%)	F-ratio$_{(2, 9)}$	P	SE
Upland	1	NPK	2.69 ± 0.006	2.84	2.88	2.85	–	0.3	0.732	0.03
		OM	2.72 ± 0.016	2.83	2.69	2.73	0.11	4.1	0.050	0.04
	2	NPK	2.99 ± 0.011	3.06	3.05	2.97	–	1.1	0.383	0.05
		OM	2.97 ± 0.006	3.15	2.96	2.96	0.09	13.1	0.002	0.03
	3	NPK	3.07 ± 0.029	3.15	2.85	3.05	0.08	42.4	<0.001	0.02
		OM	3.07 ± 0.017	3.43	3.08	3.03	0.15	21.5	<0.001	0.05
Paddy	1	NPK	2.96 ± 0.013	3.65	2.96	2.69	0.12	189.5	<0.001	0.04
		MNPK	2.89 ± 0.031	3.72	3.00	2.79	0.12	171.9	<0.001	0.04
	2	NPK	3.08 ± 0.028	3.19	3.01	2.80	0.05	151.3	<0.001	0.02
		MNPK	3.06 ± 0.011	3.29	3.00	3.01	0.05	96.4	<0.001	0.02
	3	NPK	2.79 ± 0.041	3.44	3.04	2.61	0.09	202.1	<0.001	0.03
		MNPK	2.69 ± 0.043	3.54	3.20	2.94	0.08	157.5	<0.001	0.02

Data of bulk soil are shown as means ± standard error ($n = 4$). F, P and SE (standard error) values represent the results of a one-way ANOVA (degrees of freedom (d.f.) = 2 for the treatment and d.f. = 9 for the error) for Q_{10} among different sizes of soil particles. LSD is the least significant difference; it is significant at $P < 0.05$. NPK, inorganic fertilizer only; OM, organic manure (cow manure); MNPK, organic manure (pig manure) plus NPK.

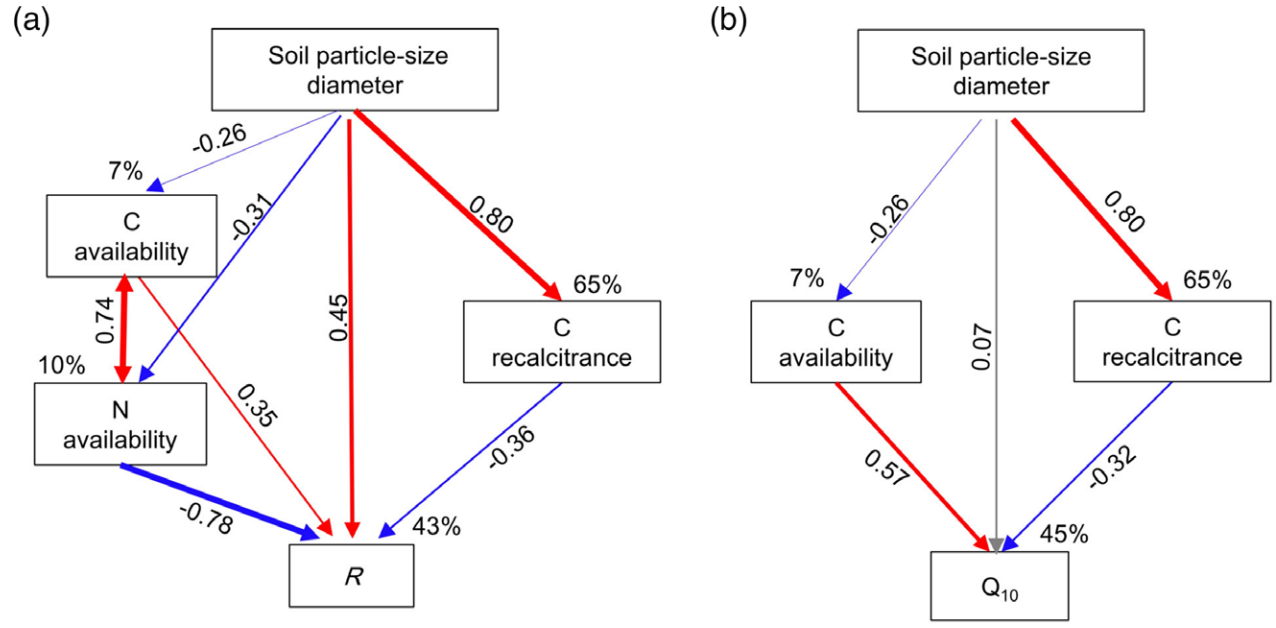

Figure 1 Structural equation models of the factors that affect (a) rates of soil C decomposition (R) at 20°C and (b) Q_{10} values. Soil C and N availability are represented by dissolved organic carbon and total dissolved nitrogen contents, respectively. Soil C recalcitrance is indicated by the soil C/N ratios. The final models fitted the data well: (a) $\chi^2 = 2.7$, d.f. = 2, χ^2/d.f. = 1.3, $P = 0.263$, AIC = 38.7, RMSEA = 0.047, for the C decomposition model and (b) $\chi^2 = 0.2$, d.f. = 1, χ^2/d.f. = 0.2, $P = 0.655$, AIC = 26.2, RMSEA = 0 for the Q_{10} model. Numbers close to the arrows are the standardized path coefficients. Double-headed arrows represent covariance between related variables. Single-headed arrows indicate the hypothesized direction of causation. Red arrows indicate significant positive relations, and blue arrows indicate significant negative relations ($P < 0.05$). Grey arrows indicate non-significant relations ($P > 0.05$). The width of the arrows indicates the strength of the relations (according to the absolute value of the standardized path coefficient). The percentages close to the endogenous variables indicate the variance explained by the model (R^2). d.f., degrees of freedom.

Discussion

Possible mechanisms of C decomposition in different soil particles: rates

Our study indicated that C in the coarse soil particles (sand fraction) had larger rates of decomposition than in fine particles (silt and clay fractions) for all the cropland soils (Table 4). This accords with previous studies (Bimüller *et al*., 2014; Ding *et al*., 2014). It is clear that this pattern is due mainly to the capacity and strength of sorption of SOM by the soil mineral surface that increases with decreasing particle size (von Lützow *et al*., 2007). This is why soil particle-size diameter (representing soil mineral sorption) shows a

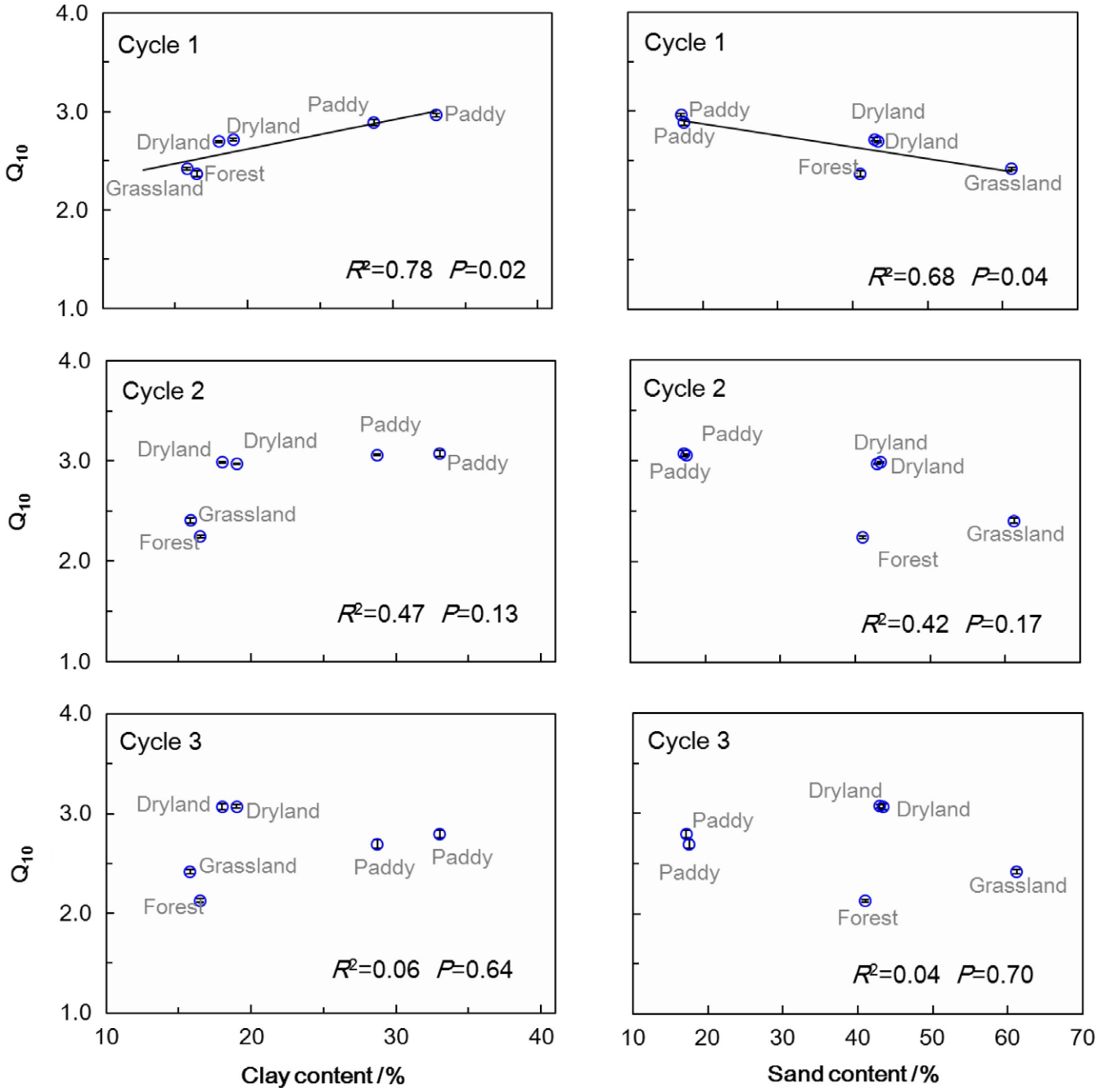

Figure 2 Relation between the Q_{10} of bulk soil and (a) clay and (b) sand contents. In each panel, the points denote the bulk soils of the paddy, upland, grassland and forest. Data for grassland and forest soils were from Ding *et al*. (2014). We did not incubate bulk soil for grassland and the Q_{10} values for grassland soil were substituted by the weighted average of their three particle-size fractions. The error bars represent the standard errors ($n = 4$).

dominating direct positive effect on *R* (Figure 1a). Moreover, soil particle size had an indirect positive effect through the availability of N (Figure 1a). Specifically, larger soil particles had smaller N availability, which conversely led to larger *R* (Figure 1a). The negative correlation between N availability and soil particle size (Figure 1a, Table S3, Supporting Information) was also observed by Bimüller *et al*. (2014). Stimulation of *R* by small N availability was supported by the 'microbial nitrogen mining' hypothesis; that is, when N availability is limited, soil microbes will accelerate decomposition of organic matter to acquire N (Craine *et al*., 2007).

On the contrary, soil particle size had indirect negative effects on *R* through C availability and recalcitrance (Figure 1a). Specifically, finer soil particles had larger C availability, which would lead to larger *R* (Figure 1a). Large C availability can boost the number

and activity of microorganisms, thereby accelerating the rates of the enzyme-catalysed processes (Davidson & Janssens, 2006). On the other hand, fine soil particles had low C recalcitrance, which would in turn lead to large R (Figure 1a). Low C recalcitrance in fine soil particles could be related to the loss of aromatic C with decreasing particle size (von Lützow *et al*., 2007). The negative relation between R and C recalcitrance indicated that it was reasonable to use the C/N ratio to indicate C recalcitrance. However, the total negative effects through C availability and recalcitrance of soil particle size on R were smaller than the total positive effect through mineral sorption and N availability. Therefore, the ultimate outcome was a positive effect: that C in larger soil particles had faster rates of decomposition (Figure S4, Supporting Information).

Although soil minerals in fine particles had larger sorption capacity for SOM than those in coarse particles, the former generally had more available C, as indicated by DOC (Table 2). This was not countered because fine particles (i.e. clay) contained more microorganisms than coarse particles, as indicated by MBC (especially for the upland soils, Table 2). Thus, the former accumulated a larger number of microbial products (e.g. diaminopimelic acids, amino sugars, teichoic acids, and so on) that were small and dissolved (available) molecules (Christensen, 2001). Although minerals in fine particles have large sorption capacity, their sorption sites were relatively limited and the remaining amount of available C was still larger compared with the coarse particles. Furthermore, ultrasonic dispersion during the soil particle-size fractionation destroyed soil aggregates, and the C occluded within the aggregates might become available and be relocated into the clay fraction.

Possible mechanisms of C decomposition in different particles: Q_{10}

Our first hypothesis was proved by the fact that the clay fraction (fine particle) had a larger Q_{10} than the silt and sand fractions (coarse particles) for most soils we investigated ($P < 0.05$, Table 5). The only exception was the upland soil under NPK treatment in the first ($P = 0.732$) and second cycles ($P = 0.383$). The P value >0.05 in the second cycle could result from the large standard error for Q_{10} of the sand fraction (Table 5). The Q_{10} trend across soil particles agreed well with our previous findings in grassland and forest soils (Ding *et al*., 2014). Moreover, Arevalo *et al*. (2012) found that medium and fine particles had larger Q_{10} values than coarse particles for most of their investigated soils, although they classified fine, medium and coarse particles as <53 μm, 53–250 μm and 250–2000 μm, respectively. Thus, we suggest that our finding of smaller C decomposition in finer soil particles that are more sensitive to temperature is probably a widespread phenomenon.

The mechanisms behind this phenomenon can be observed in Figures 1(b) and S4 (Supporting Information). Above all, soil particle size had an indirect negative effect on Q_{10} through soil C availability (Figure 1b). Specifically, the large-sized particles had smaller C availability that resulted in a smaller Q_{10}. The positive relation between Q_{10} and C availability was in accordance with previous studies (Gershenson *et al*., 2009; Fissore *et al*., 2013). When C availability is not limited, Q_{10} will be expressed as the intrinsic temperature sensitivity (inherent kinetic properties of the reaction rate, not affected by the environment), whereas when C availability is small, the temperature sensitivity of K_m (the Michaelis–Menten constant) and V_{max} (the maximum reaction rate at a given temperature) can neutralize each other, creating an apparent small Q_{10} value (Davidson & Janssens, 2006; Davidson *et al*., 2006).

Soil particle size had an indirect negative effect on Q_{10} through C recalcitrance (Figure 1b). Specifically, larger particles had more recalcitrant C, which resulted in a smaller Q_{10}. The negative relation between Q_{10} and C recalcitrance conflicted with the kinetic theory (Bosatta & Ågren, 1999). In our study, we did not measure the molecular-level characteristics of the organic matter associated with different soil particles. Although we found a negative relation between C decomposition and C recalcitrance indicted by C/N (Figure 1a), the C/N ratio might be a limited and imperfect indicator of C recalcitrance when investigating soil mineral fractions.

Soil particle size, however, had no direct effect on Q_{10} (Figure 1b), although C decomposition in finer soil particles generally had larger Q_{10} values (Table 5) and Q_{10} was significantly correlated with soil particle-size diameter (Table S3, Supporting Information). Considering that soil mineral sorption capacity and strength may play a critical role in the direct effect of soil particle size (as assumed above), the adsorption of SOM by soil minerals (i.e. chemical protection) could have no effect on the temperature sensitivity of C decomposition. This view was partly supported by Moni *et al*. (2015), who found that C decomposition in organic soil (no or slight adsorption of SOM by soil minerals) and mineral soil (strong adsorption) had similar Q_{10} values. Contrastingly, Tang & Riley (2015) found that increasing the capacity and activation energy of mineral sorption increased Q_{10} through model simulation. Davidson & Janssens (2006) assumed that mineral sorption with SOM (i.e. chemical protection) would inhibit Q_{10}, given that mineral sorption can reduce substrate availability. There is still no consensus among scientists on this topic because mineral sorption itself can respond to changing temperature and different types of chemical bonds have different responses (Conant *et al*., 2011).

Effect of soil texture on Q_{10}

Our study indicated that Q_{10} was positively correlated with the clay content ($r = 0.88$, $P = 0.02$) and negatively correlated with the sand content ($r = 0.91$, $P = 0.04$, Figure 2) during the first cycle, which seems to support our second hypothesis. However, the dependence of Q_{10} on soil texture decreased and even disappeared during the second and third cycles. Our study suggested that C decomposition in fine-textured soil could be more sensitive to global warming than in coarse-textured soil in the short term, whereas in the long term the pattern might change. This suggestion was partly supported by Zhou *et al*. (2014), who found that Q_{10} values were larger in a silty clay soil (clay content, 34.9%) than in a clay loam soil (clay, 29.0%) and a sandy loam soil (clay, 17.6%) during early incubation, but in late incubation the clay loam soil had the largest Q_{10}. Likewise,

Froseth & Bleken (2015) showed that Q_{10} was larger in a silty clay loam than in a sandy loam soil during 0–15 days incubation, but the trend was reversed after a longer incubation period (15–142 days).

The dependence of Q_{10} on soil texture and its independence have both been reported previously; they might be related to different durations of incubation. Ding *et al*. (2016) found that Q_{10} was negatively correlated with sand content after 10 days of pre-incubation for soils at 156 grassland sites across the Tibetan Plateau. However, Wei *et al*. (2014) did not observe the effect of clay content on Q_{10} after 30 days of pre-incubation with artificial soils. Hamdi *et al*. (2013) did not note a significant relation between Q_{10} and soil texture in a meta-analysis using a dataset that included both short- and long-term incubation.

Although C decomposition in finer soil particles is more sensitive to temperature, there was not a long-lasting relation between the Q_{10} of bulk soil and soil texture (Figure 2). A possible explanation is that occlusion of SOM within aggregates (i.e. physical protection) in bulk soil (not existing in soil particles) changed the expected long-lasting outcome. The amount of newly-formed soil aggregates and physical protection for C might increase with incubation time. Fine-textured soil, with large clay content and large SOM content, improved aggregate formation and physical protection more (Wagner *et al*., 2007), which might obscure the temperature sensitivity by reducing substrate availability (Davidson & Janssens, 2006). Therefore, fine-textured soil did not have larger Q_{10} than coarse-textured soil in the long term.

Q_{10}: labile C compared with stable C pool

Differently sized particle fractions represent C pools with different stabilities (von Lützow *et al*., 2007). The clay fraction generally had the slowest rate of C decomposition among the three particle fractions (Table 4), and the C in this fraction has been reported to be stable and have a long mean residence time (von Lützow *et al*., 2007). However, this pool had the largest Q_{10} among the three particle-size fractions (Table 5). Moreover, the Q_{10} values had a significant negative correlation with the rates of C decomposition (Table S3, Supporting Information). This clearly indicates that the passive C pool (slow turnover) would be more sensitive to climate warming than the labile C pool (fast turnover). This finding is in accord with many previous studies (e.g. Craine *et al*. 2010). The passive C pool usually occupies a large part of the total soil C. For instance, the clay fraction contained a larger amount of C than the sand fraction in most soil types (Table 1). Therefore, the long-term response of global soil C stocks to climate warming is more severe than the current predictions using the same Q_{10} (usually equal to 2) for all C pools.

Conclusions

The temperature sensitivity of C decomposition was generally larger in fine soil particles than in coarse particles for cropland soils regardless of water and fertilizer management, which was the same as the results of our previous study of grassland and forest soils. We suggest that C decomposition in fine soil particles is more sensitive to climate warming than in coarse particles, and that this is a widespread phenomenon. However, this phenomenon did not bring a long-lasting relation between the Q_{10} of bulk soil and soil texture. The results suggested that the C decomposition in fine-textured soil was more sensitive to global warming than that in coarse-textured soil in the short term, whereas in the long term the pattern might change.

Supporting Information

The following supporting information is available in the online version of this article:

Figure S1. Schematic diagram of the time sequence for the three temperature-change cycles.

Figure S2. Relation between rates of carbon decomposition in particle-size fractions of upland soil and incubation temperature.

Figure S3. Relation between rates of carbon decomposition in particle-size fractions of paddy soil and incubation temperature.

Figure S4. Schematic diagram of the mechanisms of different rates of C decomposition in differently sized soil particles.

Figure S5. Schematic diagram of the mechanisms of different Q_{10} in differently sized soil particles.

Table S1. Results of structural equation modelling of the factors influencing rates of soil C decomposition as illustrated in Figure 1(a).

Table S2. Results of structural equation modelling of the factors influencing Q_{10} as illustrated in Figure 1(b).

Table S3. Pearson correlation coefficients between soil variables.

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References

Arevalo, C.B.M., Chang, S.X., Bhatti, J.S. & Sidders, D. 2012. Mineralization potential and temperature sensitivity of soil organic carbon under different land uses in the parkland region of Alberta, Canada. *Soil Science Society of America Journal*, **76**, 241–251.

Bimüller, C., Mueller, C.W., von Lützow, M., Kreyling, O., Kölbl, A., Haug, S. *et al.* 2014. Decoupled carbon and nitrogen mineralization in soil particle size fractions of a forest topsoil. *Soil Biology & Biochemistry*, **78**, 263–273.

Bond–Lamberty, B. & Thomson, A. 2010. Temperature-associated increases in the global soil respiration record. *Nature*, **464**, 579–582.

Bosatta, E. & Ågren, G. 1999. Soil organic matter quality interpreted thermodynamically. *Soil Biology & Biochemistry*, **31**, 1889–1891.

Chen, X., Tang, J., Jiang, L., Li, B., Chen, J. & Fang, C. 2010. Evaluating the impacts of incubation procedures on estimated Q_{10} values of soil respiration. *Soil Biology & Biochemistry*, **42**, 2282–2288.

Christensen, B.T. 2001. Physical fractionation of soil and structural and functional complexity in organic matter turnover. *European Journal of Soil Science*, **52**, 345–353.

Conant, R.T., Ryan, M.G., Ågren, G.I., Birge, H.E., Davidson, E.A., Eliasson, P.E. *et al.* 2011. Temperature and soil organic matter decomposition rates – synthesis of current knowledge and a way forward. *Global Change Biology*, **17**, 3392–3404.

Craine, J.M., Morrow, C. & Fierer, N. 2007. Microbial nitrogen limitation increases decomposition. *Ecology*, **88**, 2105–2113.

Craine, J.M., Fierer, N. & McLauchlan, K.K. 2010. Widespread coupling between the rate and temperature sensitivity of organic matter decay. *Nature Geoscience*, **3**, 854–857.

Davidson, E.A. & Janssens, I.A. 2006. Temperature sensitivity of soil carbon decomposition and feedbacks to climate change. *Nature*, **440**, 165–173.

Davidson, E.A., Janssens, I.A. & Luo, Y.Q. 2006. On the variability of respiration in terrestrial ecosystems: moving beyond Q_{10}. *Global Change Biology*, **12**, 154–164.

Ding, F., Huang, Y., Sun, W., Jiang, G. & Chen, Y. 2014. Decomposition of organic carbon in fine soil particles is likely more sensitive to warming than in coarse particles: an incubation study with temperate grassland and forest soils in Northern China. *PLoS ONE*, **9**, e95348.

Ding, J., Chen, L., Zhang, B., Liu, L., Yang, G., Fang, K. *et al.* 2016. Linking temperature sensitivity of soil CO_2 release to substrate, environmental, and microbial properties across alpine ecosystems. *Global Biogeochemical Cycles*, **30**, 1310–1323.

Fang, C.M., Smith, P., Moncrieff, J.B. & Smith, J.U. 2005. Similar response of labile and resistant soil organic matter pools to changes in temperature. *Nature*, **433**, 57–59.

Fissore, C., Giardina, C.P. & Kolka, R.K. 2013. Reduced substrate supply limits the temperature response of soil organic carbon decomposition. *Soil Biology & Biochemistry*, **67**, 306–311.

Froseth, R.B. & Bleken, M.A. 2015. Effect of low temperature and soil type on the decomposition rate of soil organic carbon and clover leaves, and related priming effect. *Soil Biology & Biochemistry*, **80**, 156–166.

Gershenson, A., Bader, N.E. & Cheng, W. 2009. Effects of substrate availability on the temperature sensitivity of soil organic matter decomposition. *Global Change Biology*, **15**, 176–183.

Grace, J. 2006. *Structural Equation Modelling and Natural Systems*. Cambridge University Press, Cambridge.

Hamdi, S., Moyano, F., Sall, S., Bernoux, M. & Chevallier, T. 2013. Synthesis analysis of the temperature sensitivity of soil respiration from laboratory studies in relation to incubation methods and soil conditions. *Soil Biology & Biochemistry*, **58**, 115–126.

von Lützow, M., Kögel-Knabner, I., Ekschmitt, K., Flessa, H., Guggenberger, G., Matzner, E. *et al.* 2007. SOM fractionation methods: relevance to functional pools and to stabilization mechanisms. *Soil Biology & Biochemistry*, **39**, 2183–2207.

Moni, C., Lerch, T.Z., de Zarruk, K.K., Strand, L.T., Forte, C., Certini, G. *et al.* 2015. Temperature response of soil organic matter mineralisation in arctic soil profiles. *Soil Biology & Biochemistry*, **88**, 236–246.

Nelson, P.N., Dictor, M.C. & Soulas, G. 1994. Availability of organic carbon in soluble and particle-size fractions from a soil profile. *Soil Biology & Biochemistry*, **26**, 1549–1555.

Ouyang, W., Shan, Y., Hao, F. & Lin, C. 2014. Differences in soil organic carbon dynamics in paddy fields and drylands in Northeast China using the CENTURY model. *Agriculture, Ecosystems & Environment*, **194**, 38–47.

Schlesinger, W.H. & Andrews, J.A. 2000. Soil respiration and the global carbon cycle. *Biogeochemistry*, **48**, 7–20.

Schulten, H.R. & Leinweber, P. 1991. Influence of long-term fertilization with farmyard manure on soil organic matter: characteristics of particle-size fractions. *Biology and Fertility of Soils*, **12**, 81–88.

Shirazi, M.A. & Boersma, L. 1984. A unifying quantitative analysis of soil texture. *Soil Science Society of America Journal*, **48**, 142–147.

Soil Survery Staff 1999. *Soil Taxonomy. A Basic System of Soil Classification for Making and Interpreting Soil Surveys*. Second Edition. Agricultural Handbook 436. Soil Conservation Service, US Department of Agriculture, Washington, DC, 163–270.

Sollins, P., Homann, P. & Caldwell, B.A. 1996. Stabilization and destabilization of soil organic matter: mechanisms and controls. *Geoderma*, **74**, 65–105.

Tang, J. & Riley, W.J. 2015. Weaker soil carbon-climate feedbacks resulting from microbial and abiotic interactions. *Nature Climate Change*, **5**, 56–60.

Tian, G., Kang, B.T. & Brussaard, L. 1992. Biological effects of plant residues with contrasting chemical compositions under humid tropical conditions – decomposition and nutrient release. *Soil Biology & Biochemistry*, **24**, 1051–1060.

Tiessen, H. & Stewart, J.W.B. 1983. Particle-size fractions and their use in studies of soil organic-matter: II. Cultivation effects on organic-matter composition in size fractions. *Soil Science Society of America Journal*, **47**, 509–514.

Wagner, S., Cattle, S.R. & Scholten, T. 2007. Soil-aggregate formation as influenced by clay content and organic-matter amendment. *Journal of Plant Nutrition and Soil Science*, **170**, 173–180.

Wang, W., Dalal, R., Moody, P. & Smith, C. 2003. Relationships of soil respiration to microbial biomass, substrate availability and clay content. *Soil Biology & Biochemistry*, **35**, 273–284.

Wei, H., Guenet, B., Vicca, S., Nunan, N., Asard, H., AbdElgawad, H. *et al.* 2014. High clay content accelerates the decomposition of fresh organic matter in artificial soils. *Soil Biology & Biochemistry*, **77**, 100–108.

Zhou, P., Li, Y., Ren, X.E., Xiao, H.A., Tong, C., Ge, T. *et al.* 2014. Organic carbon mineralization responses to temperature increases in subtropical paddy soils. *Journal of Soils Sediments*, **14**, 1–14.