Vertical and seasonal variations in soil CO₂ production in a 55-year-old oriental arborvitae (*Platycladus orientalis*) plantation in China

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ABSTRACT

The aim of this work was to explore the vertical and seasonal variations of CO₂ within the soil profile based on Fick's law of diffusion in an oriental arborvitae plantation. We continuously measured the soil CO₂ concentration profile using CO₂ sensors buried at different depths in a coniferous forest in northern China and calculated the CO₂ flux based on the profile measurements using a dynamic model. The diurnal pattern of CO₂ concentration and flux fluctuated during the day and varied less at night. The CO₂ profile had a vertical gradient, with the highest concentrations in the deepest soil layers. The CO₂ flux had a clear seasonal pattern with a maximum in summer and a minimum in winter. The contributions of the H, A, B, and C horizons to the total CO₂ flux were 75.38, 13.52, 7.61, and 3.49%, respectively. Q₁₀ was 2.668, 4.469, 1.175, and 3.333 in the H, A, B, and C horizons, respectively. The CO₂ flux determined from the concentration profiles agreed well with the CO₂ flux measured by open dynamic chambers, suggesting that this model could be used to successfully measure soil CO₂ emissions and to describe the processes underlying CO₂ efflux.

Keywords: *Platycladus orientalis* plantation, soil CO₂ profile, flux, temperature, water content

RÉSUMÉ

Nous avons voulu par ce projet étudier les variations verticales et saisonnières du CO₂ dans le profil pédologique d'après la Loi de la diffusion de Fick dans une plantation de thuya oriental. Nous avons mesuré en continu le profil de concentration de CO₂ dans le sol au moyen de détecteurs de CO₂ enfouis à différentes profondeurs dans le sol d'une forêt de conifères du nord de la Chine et calculé le flux de CO₂ à partir des mesures effectuées le long de ce profil à l'aide d'un modèle dynamique. Le modèle diurne de la concentration de CO₂ ainsi que son flux variaient davantage au cours de la journée que durant la nuit. Le profil de CO₂ présentait un gradient vertical dont les plus fortes concentrations se retrouvaient dans les couches les plus profondes du sol. Le flux de CO₂ suivait un modèle nettement saisonnier, le maximum étant atteint en été et le minimum au cours de l'hiver. Les apports en provenance des horizons H, A, B et C au flux total de CO₂ étaient respectivement de 75,38; 13,52; 7,61 et 3,49 %. Le Q₁₀ se chiffrait à 2,668; 4,469; 1,175 et 3,333 dans les horizons H, A, B et C respectivement. Le flux de CO₂ établi à partir des profils de concentration était conforme aux flux de CO₂ mesurés en chambre à flux dynamique, laissant entendre que le modèle pourrait être utilisé pour mesurer correctement les émissions de CO₂ en provenance du sol et pour décrire les processus d'échappement du CO₂.

Mots clés : plantation de *Platycladus orientalis*, profil de CO₂ du sol, flux, température, teneur en eau

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Introduction

Increasing the area of forested land via reforestation is a major strategy for mitigating the effects of carbon emissions. Planted forests in China have primarily contributed to increasing the accumulation of carbon since the 1970s (Fang *et al.* 2001). Oriental arborvitae (*Platycladus orientalis* (Linn. Franco) is an important species for reforestation in temperate areas. However, planted forests are more vulnerable than natural forests to environmental stress (Chazdon 2008, Li *et al.* 2016) and their ability to sequester CO₂ is uncertain. The role of soil in China's terrestrial carbon balance is a large source of this uncertainty (Piao *et al.* 2009). Soil respiration has been determined in most districts of China (Tian *et al.* 2010, Yu *et al.* 2011, Pang *et al.* 2012), but due to the limitations of measuring techniques, vertical and seasonal variations of soil CO₂ in planted forests is limited.

Soil CO₂ flux, or soil respiration, plays an important role in the global carbon cycle, constituting approximately 75% of total ecosystem respiration (Law *et al.* 2001). The potential and controversial role of soil CO₂ efflux in accelerating global warming has led to increased attention (Rodeghiero and Cescatti 2005, Vargas and Allen 2008, Li *et al.* 2016). CO₂ released from the soil to the atmosphere could possibly contribute to a positive feedback between increasing temperatures and enhanced CO₂ efflux and then to an acceleration of global warming (Rodeghiero and Cescatti 2005, Tucker and Reed 2016). Improved accuracy in estimations of the carbon budget of terrestrial ecosystems requires a reduction of the uncertainties of measurements. CO₂ efflux is usually measured in chambers on the soil surface, but these chambers are unable to provide information about the vertical distribution of CO₂ sources and individual contributions at certain depths, which are important for understanding the mechanisms of soil carbon transfer. Also, a reliable and robust automated chamber for field measurements is not commercially available (Tang *et al.* 2003, Pumpanen *et al.* 2008).

An early method for studying the soil CO₂ profile and diffusion was the periodic extraction of gas samples from different depths (Davidson and Trumbore 1995). This method can provide information for CO₂ production at different depths but cannot provide *in situ* and continuous data for CO₂ flux (Tang *et al.* 2003), and does not have a high temporal resolution over long time periods because it is destructive and quite laborious.

Biological activity in different soil layers with high temporal resolution for long periods would provide better insight into the processes underlying CO₂ fluxes than would the conventional closed chambers (Pumpanen *et al.* 2008). Several studies have estimated soil CO₂ flux from vertical CO₂ concentration gradients (Pihlatie *et al.* 2007, Delogu *et al.* 2016). This method can continuously and automatically determine CO₂ flux at various temporal scales with minimal disturbance upon installation and thus can provide a good opportunity for studying the processes underlying CO₂ fluxes *in situ* with fewer disturbances.

In this study, we used permanently installed CO₂ probes to automatically and continuously determine the soil CO₂ profile in a typical oriental arborvitae plantation in northern China. The CO₂ fluxes were calculated from the concentration profile using a dynamic model. We used a gradient method to determine the relationship between CO₂ flux, tem-

perature and moisture content, and validated it using closed chambers. The aims of this research were: (1) to determine the distribution of CO₂ concentrations in the soil profile of a typical planted forest; (2) to examine the diurnal and seasonal changes in CO₂ flux in the profile; (3) to estimate the relative contributions of CO₂ sources to total respiration at various depths; and, (4) to understand the response of soil respiration to soil temperature and moisture content.

Materials and Methods

Site description

All measurements were performed in a 55-year-old oriental arborvitae plantation on Jiufeng Mountain (39° 54' N, 116° 28' E; 145 m.a.s.l.), Beijing, China. The area has a warm temperate, semi-humid continental monsoon climate. Mean annual temperature is 12.2 °C. The highest temperature recorded in this area was 41.9 °C, and the lowest was -27.4 °C. Mean precipitation is 650–720 mm, 60%–80% of which occurs from June to September. The soil is a shallow (0.3–1.0 m, 0.55 m average) loessial gravelly loam with a 30% coarse (> 2 mm) fraction. Soil pH is 6.49 ± 0.24, and the contains 13.42 ± 1.68 g kg⁻¹ organic matter, 1.67 ± 0.22 g kg⁻¹ total nitrogen, 4.56 ± 0.29 mg kg⁻¹ available phosphorus, and 143.75 ± 17.16 mg kg⁻¹ available potassium.

The trees have an average height of 10.7 m and a mean diameter at breast height of 20.9 cm. Stand density is 1176 stems per hectare. A sparse deciduous understory consists of such species as the chaste tree (*Vitex negundo* var. *cannabifolia* L.), morrow honeysuckle (*Grewia biloba* var. *parviflora* L.), and wild jujube (*Ziziphus jujuba* var. *spinosa* L.).

Measurements of CO₂ concentrations

Soil CO₂ concentrations were measured *in situ* using Vaisala GMP343 diffusion CO₂ sensors (Vaisala Corp., Vantaa, Finland) with a range of 0–20 000 µmol mol⁻¹, based on Vaisala's advanced CARBOCAP Single-Beam Dual Wavelength non-dispersive infra-red technique.

The GMP343 sensors were installed in a vertical face of a soil pit excavated at the site while maintaining the separation of soil layers. Four probes were installed at depths of 0.03, 0.10, 0.23, and 0.40 m in one vertical profile (Fig. 1). Smaller holes (approximately 0.10 m deep) were cut into the face of the pit where the sensors were placed so that the slot in each sensor head pointed downward to avoid trapping water inside the protective cap of the sensor. After installation, the soil layers were replaced in the same order. There were three vertical profiles for three replicates.

Soil temperature (*T*) and volumetric water content (*θ*) were monitored at one-hour intervals using intelligent rod sensors (TRIM-PICO3232, IMKO GmbH) installed at the same depths as the CO₂ sensors. *T* and *θ* were used to calculate the diffusivity of each layer. A schematic presentation of the setup is presented in Fig. 1. Hourly measurements of CO₂ concentration, *T*, and *θ* were recorded in the profile with a DT-80 dataTaker (Thermo Fisher Scientific Australia Pty Ltd, Scoresby, Australia).

Measurement of CO₂ flux

To validate the CO₂ gradient method, CO₂ flux was measured by an automated system with a Li-8100-104 long-term chamber (LICOR Environmental, Lincoln, Nebraska USA). The

Fig. 1. Schematic presentation of sensor installation. Arrows on the right indicate the thickness of the layers; left arrows indicate installation depth of the sensors.

automated chamber on a PVC collar had been used for monitoring CO₂ efflux at one-hour intervals during 00:00 to 24:00 hours in the same position. The PVC collar (11.4 cm height; 21.3 cm diameter) was recessed approximately 8 cm into the forest floor. Five other PVC collars were randomly and permanently placed in the measurement site to determine the spatial heterogeneity of the CO₂ efflux. To avoid disturbing the continuous CO₂ efflux measurement in the pit, the CO₂ concentration profile was constructed 6 m from the chamber. The ambient CO₂ concentration in the chamber was measured at the soil surface to calculate the diffusion.

CO₂ gradient method

The CO₂ flux was calculated with a dynamic model based on Fick's law of diffusion (Pumpanen *et al.* 2008). In this model, soil is described as a layered structure divided into different horizons. The horizons were humus (H horizon), eluvial (A horizon), illuvial (B horizon), and parental material (C horizon). Fig. 1.

The movement of CO₂ from the soil to the atmosphere is estimated by diffusion, which is dependent on the porosity, water content, and thickness of the soil layers and on the concentration gradient between the layers. The soil at the study site is shallow, averaging ~0.55 m over solid bedrock, so the vertical movement of air in the soil and the storage of CO₂ in the air spaces are very low (Fang and Moncrieff 1999).

The CO₂ flux between the A and H horizons was calculated as:

$$(1) \quad F_{A-H} = -D_A \frac{C_H - C_A}{(L_H + L_A) / 2}$$

Where: F_{A-H} is the flux from the A to the H horizon (g CO₂ m⁻² s⁻¹), D_A the diffusion coefficient of CO₂ in the A horizon (m² s⁻¹), C_H and C_A the CO₂ concentrations (g CO₂ m⁻³) in H

and A horizons, respectively, and L_H and L_A the thicknesses (m) of H and A horizons, respectively. The transport was calculated between the H horizon and the atmosphere (ATM), B to A horizons, and C to B horizons. It was assumed that transport between the H horizon and the atmosphere was similar to transport between horizons (Pihlatie *et al.*, 2007), and the distance $(L_H + L_{ATM})/2$ was simply the thickness of the H horizon divided by 2.

We calculated the diffusion coefficient of CO₂ in the soil, D , from the diffusion coefficients in free air (D_0) (m² s⁻¹) (Troeh *et al.*, 1982):

$$(2) \quad \frac{D}{D_0} = \left\{ \frac{Eg - u}{1 - u} \right\}^n$$

Where: Eg is the soil air-filled porosity (m³ m⁻³), and u and n are empirical parameters of soil tortuosity (Glinski and Stepiński 1985). D was estimated separately for each horizon. Eg was obtained by subtracting θ from the total porosity (E_0):

$$(3) \quad Eg = E_0 - \theta$$

E_0 was obtained from the soil-water retention curves estimated for the soil layers (Mecke and Ilvesniemi 1999). Soil-water retention curves for each soil horizon were determined from soil samples collected with steel cylinders (0.05 m diameter; 0.11 m length) from five excavated pits at the study site. The values of the parameters used in Eqs. 1-3 are presented in Table 1.

The effect of temperature on D_0 calculated as (Armstrong 1979):

$$(4) \quad \log(D_0) = 1.9975 \log(T) - 9.7273$$

Where: T is the temperature (K).

Table 1. Parameters used in the calculation of CO₂ transport functions [Eq.(1-3)]

Soil horizon	<i>u</i>	<i>h</i>	<i>L</i> (m)	<i>Eo</i> (m ³ m ⁻³)
H-horizon	0	1.1	0.050	0.675
A-horizon	0	1.4	0.100	0.632
B-horizon	0	1.4	0.150	0.593
C-horizon	0	1.4	0.200	0.566

u and *h* are empirical parameters obtained; *L* is the thickness of soil layer; *Eo* is the total porosity of the soil.

Changes in θ occurred rapidly during rain events, and it was assumed that the movement of CO₂ between the layers also pushed the water, thereby replacing the air in the soil pores (Pumpanen *et al.* 2003, Pumpanen *et al.* 2008). For example, the CO₂ flux from the A to the H horizon (J_{AHP}) caused by changes in the air-filled porosity of the A horizon was expressed as:

$$(5) \quad J_{AHP} = \frac{[E_{gA}(t_i) - E_{gA}(t_{i+1})]L_A C_{gA}(t_i)}{t_{i+1} - t_i}$$

Where: $E_{gA}(t_i)$ is the air-filled porosity (m³ m⁻³) of the A horizon at time t_i , and $C_{gA}(t_i)$ is the CO₂ concentration (g CO₂ m⁻³) in the A horizon at time t_i .

The CO₂ flux was calculated based on the model using hourly values of soil and atmospheric CO₂ concentrations (g CO₂ m⁻³), T (°C), and θ (m³ m⁻³). Soil respiration in each layer was obtained with a CO₂ mass balance equation. As an example, the respiration in the A horizon (R_A) was:

$$(6) \quad R_A = J_{A-H} - J_{B-A} + J_{AHP} + (C_A(t_{i+1}) - C_A(t_i))E_{gA}$$

The well-known Arrhenius and Q_{10} (temperature sensitivity coefficient) functions were used to describe the responses of respiration to T in each layer during the measurement period (Lloyd and Taylor 1994):

$$(7) \quad R_s = ae^{bT}$$

Where: R_s is soil respiration (g CO₂ m⁻² s⁻¹), and a and b are the fitting parameters. Q_{10} was estimated by following (Boone *et al.* 1998):

$$(8) \quad Q_{10} = e^{10b}$$

Statistical analysis

The dynamic model used for calculating the transport of CO₂ in soil and the respiration in the various layers was implemented with EXCEL (2010, Washington, USA). Statistical analyses and the response of soil respiration to T and θ in individual layers were performed with SPSS (V 17.0, Chicago, USA).

Results

Diurnal variation of CO₂ concentration in soil profile

The diurnal variation of CO₂ concentration, T , and θ in the soil profile showed a vertical gradient (Fig. 2). The higher CO₂ concentrations and θ increased with increasing depth. T was

Fig. 2. Diurnal variation of CO₂ concentration (a), temperature (b), volumetric water content (c).

higher in the topsoil than in the deeper soil in summer while lower in winter. The CO₂ concentrations peaked near 10:00 hours, and T in the H and A horizons peaked near 19:00, but the B and C horizons did not have the same trend. Moreover, θ did not significantly vary diurnally ($P > 0.05$; Fig. 2).

CO₂ concentrations in the H horizon and the atmosphere fluctuated diurnally, as did the CO₂ flux measured with the gradient method and the chamber. A time-lag effect was found between the CO₂ concentration above the soil surface and in the soil, which were peaked at 14:00 hours and 9:00 in the H horizon and the atmosphere, respectively. The efflux peaked at 15:00 in the chamber and at 10:00 in the gradient method (Fig. 3).

Vertical distribution of CO₂ concentration in soil profile

CO₂ concentrations in the profile showed a positive vertical gradient with increasing depth. The concentrations fluctuated little in winter and early spring, but fluctuated widely in summer. Concentrations gradually increased with increasing depth. The mean concentrations were 2.13, 1.47, and 1.05 times higher in the C horizon than in the H, A, and B horizons, respectively (Fig. 4a). The variation trend of CO₂ concentration showed the same pattern as T and θ (Fig. 4a-c). For example, CO₂ concentrations increased sharply in summer, and both θ and T also rose sharply. In this study, CO₂ concentrations increased markedly after rain, and θ was highest in the C horizon and lowest in the H horizon (Fig. 4b).

Fig. 3. Diurnal pattern in CO₂ concentration above the soil surface and in the H-horizon, and CO₂ efflux measured by the gradient method and with chamber.

Fig. 4. Soil CO₂ concentration(a), volumetric water content(b), and temperature(c) at respective horizons between 19 September 2012 and 21 April 2014.

θ was highest in the C horizon than in other horizons throughout the measurement time while the trend of T was different with θ . Higher temperatures were in the lowest horizons in spring and winter, but lowest in summer and autumn. An opposite trend showed in the upper layers (Fig. 4c).

CO₂ flux in soil profile

The CO₂ flux from the concentration gradients had clear seasonal patterns with a maximum during summer and a minimum during winter (Fig. 5). Mean CO₂ fluxes were 0.65 ± 0.29 , 0.27 ± 0.24 , 0.11 ± 0.04 , and 0.03 ± 0.02 g CO₂ m⁻² h⁻¹ from the H horizon to the atmosphere, A to H horizons, B to A horizons, and C to B horizons, respectively. The flux was 0.52 ± 0.37 g CO₂ m⁻² h⁻¹ measured by the chamber. F_{H-ATM} estimated by the gradient method agreed relatively well with the fluxes measured with the automated closed chamber ($R^2=0.678$, $P<0.001$).

Soil respiration and its relationship with T and θ

The respiration determined from the CO₂ concentrations in individual soil layers with the mass balance equation (Eq. 6) showed a clear vertical distribution (Fig. 6). The contributions of the H, A, B, and C horizons to the total CO₂ flux averaged over the study period were 75.38%, 13.52%, 7.61%, and 3.49%, respectively.

CO₂ fluxes fluctuated highly in summer, corresponding to heavy rains (Fig. 5). The fluctuations were due to occasional higher fluxes from the A to the H horizon than from the H horizon to the atmosphere calculated with Eq. 6. The seasonal pattern of soil respiration was similar to that of T , increasing as spring progressed, peaking in summer, and declining throughout autumn to a minimum in winter.

The response of soil respiration to T was clearly seen in each soil horizon, especially the H horizon. Respiration increased with temperature, increasing sharply in the H horizon (Fig. 7a, c, e, f). Soil respiration responded to θ , especially in the H horizon, where respiration and θ were linearly correlated (Fig. 7b).

Soil respiration and T were exponentially correlated, with the highest R^2 (0.555) in the H horizon. Q_{10} calculated for the measurement period was 2.668 ($R^2 = 0.555$) for the H-horizon, 4.469 ($R^2 = 0.225$) for the A horizon, 1.175 ($R^2 = 0.199$) for the B horizon, and 3.333 ($R^2 = 0.119$) for the C-horizon (Table 2).

Discussion

Diurnal variation of soil CO₂

The diurnal variation of CO₂, temperature (T), and volumetric water content (θ) showed a vertical gradient (Fig. 2). CO₂ concentrations and θ were highest in the deepest horizons and T highest in the uppermost horizon. Similar results have been reported that, unlike higher CO₂ concentrations and moisture content with increasing depth, T was higher in the topsoil than in the deeper soil in summer (Pangintha *et al.* 2010). CO₂ concentrations peaked near 10:00 hours, and T in the H and A horizons peaked near 19:00, but the B and C horizons did not have the same trend. The diurnal variation of CO₂ in the various layers is likely due to the differential distribution of soil moisture and carbon content in the profile (Tang *et al.* 2003). Moreover, θ did not significantly vary diurnally,

in agreement with the findings reported by Tucker and Reed (2016).

CO₂ concentrations in the H horizon and the atmosphere, and the CO₂ flux measured with the gradient method and the chamber fluctuated diurnally, and the hour at which their respective peaked was different showed a time lag effect. The diurnal pattern and fluctuation may be due to soil temperatures. Temperatures at the soil surface thus had a major effect on CO₂ efflux (Pang *et al.* 2012, Delogu *et al.* 2016).

Vertical distribution of CO₂

CO₂ showed a vertical gradient and a seasonal fluctuation perhaps attributed to temperature and water content (Tang *et al.* 2003, Pingintha *et al.* 2010). They are the major abiotic factors controlling soil CO₂ concentrations, which often have a temperature-driven seasonal pattern, and summer rains can lead to increases in CO₂ concentrations (Yu *et al.* 2011, Chang *et al.* 2016). After a rainfall, CO₂ concentrations rose markedly and θ was highest in the deepest horizon. These results indicate that CO₂ concentration was strongly affected by water content, which may be due to the stimulation of biological activity following a summer rain when CO₂ fluxes increase (Lee *et al.* 2004). Rainwater could block soil pores and decrease diffusion rates, preventing CO₂ from escaping from the soil and thus increasing CO₂ concentration (Unger *et al.* 2010). In addition, temporal and spatial, both horizontal and vertical variation in soil properties, active respiration of roots and mycorrhiza, stones, and the spatial variation in quantity and quality of organic matter affect the distribution of CO₂ in soils (Vargas and Allen 2008, Li *et al.* 2016).

T and θ did not follow the same seasonal trend. These results show that the deepest horizons were the warmest during the winter because the topsoil may be frozen for several months (Pang *et al.* 2012). In contrast, the topsoil is warmer than the deeper layers in summer. Biological activity in different layers is thus highly variable seasonally, affecting CO₂ production and diffusivity.

Soil CO₂ flux in profile

The CO₂ flux had a clear seasonal pattern with a maximum during summer and a minimum during winter, and the mean CO₂ fluxes from the H horizon to the atmosphere fit well with the flux measured by the chamber ($R^2 = 0.68$; $P<0.001$). Tang *et al.* (2003) found that fluxes calculated by the gradient method agreed well with those measured by the chamber. However, Turcu *et al.* (2005) reported that CO₂ flux estimated with the gradient method was about three times larger than that measured with the chamber. This may be attributed to mass flow not considered in this model, or the gas diffusion coefficient may vary for different soils at different study areas. For calculations of CO₂ flux using the diffusion coefficient of CO₂, low and high air-filled porosities and other soil properties can all impact the effective gas diffusion coefficient in a given soil layer (Pingintha *et al.* 2010). Non-uniform vertical distributions of θ , especially under field conditions, must be considered (Tucker and Reed 2016) because they could lead to over- or underestimates of CO₂ efflux, depending on the model.

Surface CO₂ flux is found by integrating the source term with depth (Cook *et al.* 1998, Tang *et al.* 2003). CO₂ flux peaked in summer and was sensitive to fluctuations in CO₂

Fig. 5. Soil CO_2 efflux determined from CO_2 gradient method at respective horizons, CO_2 effluxes measured with automated chamber, and the rain.

Fig. 6. Soil respiration in different soil horizons determined from soil CO_2 concentration.

Fig. 7. Relationship between the one-hourly soil respiration and soil temperature (**a, c, e, g**) and the relationship between the one-hourly soil respiration and soil volumetric water content (**b, d, f, h**) in different soil horizons.

Table 2. Fit of $R_s = ae^{bT}$ to determine the relationship between one-hourly soil respiration and temperature, and $R_s = a\theta + b$ for the relationship between one-hourly soil respiration and water content.

	Soil layer	Equation	R ²	Q ₁₀
Soil temperature	H-horizon	$R_s = 0.162 e^{0.098 T}$	0.555	2.668
	A-horizon	$R_s = 0.015 e^{0.150 T}$	0.255	4.469
	B-horizon	$R_s = 0.029 e^{0.016 T}$	0.199	1.175
	C-horizon	$R_s = 0.006 e^{0.121 T}$	0.119	3.333
Soil water content	H-horizon	$R_s = 8.091 \theta - 0.624$	0.383	–
	A-horizon	$R_s = 2.854 \theta - 0.327$	0.204	–
	B-horizon	$R_s = 0.577 \theta - 0.033$	0.067	–
	C-horizon	$R_s = 0.733 \theta - 0.108$	0.136	–

concentrations and θ , especially in the H and A horizons. CO₂ flux varied seasonally in our study, gradually increasing in spring and early summer and peaked in July and August. This increase paralleled the increase in T . CO₂ flux calculated by the model depended on soil porosity and especially on T and θ . T decreased with soil depth to the C horizon in summer, but the deepest soil horizons were the warmest in winter. Soil water could block the soil pores and decrease diffusion rates. Biological activities in the various soil layers vary seasonally and can contribute to variations in CO₂ flux (Karhu *et al.* 2015, Delogu *et al.* 2016).

Soil respiration and its relationship with T and θ

The contributions of the H horizon to the total CO₂ flux over the study period averaged 75.4%. These results agree well with over 90% of soil respiratory activity concentrated in the topsoil (Pietikäinen *et al.* 1999, Pumpanen *et al.* 2008) and may be attributed to the readily available decomposable organic matter and fine roots concentrated in the surface horizons (Vargas and Allen 2008, Chang *et al.* 2016).

CO₂ and respiration fluctuated highly in summer and the seasonal pattern of soil respiration was similar to that of T , peaking in summer, likely due partly to changes in the activities of microbes and roots. The temperature sensitivity reflects both root and mycorrhizal respiration and the decomposition of labile root-derived organic material (Boone *et al.* 1998). Also, θ affects soil porosity and consequently reduces the diffusion rate of CO₂. Sharp increases in θ could disturb respiration rates. CO₂ dissolved in water may move rapidly downward. The amount of water movement in soil after a rain however, is relatively small and slow, even after a heavy rain. The movement of water in the soil profile is thus very slow, and the transport of CO₂ in dissolved water is not substantial (Lee *et al.* 2004).

The response of soil respiration to temperature was clearly seen in each soil horizon, especially the H horizon. Moreover, soil respiration increased sharply with temperature and was different in different horizons. Increase in respiration maybe due to root and microbial respiration increase with temperature, and enhanced respiratory activity of root systems increases the CO₂ flux. The trees grow rapidly in summer, and warmer temperatures and frequent rains enhance root

and microorganism activities and accelerate litter decomposition, all of which contributes to a sharply higher respiration in summer (Pang *et al.* 2012, Karhu *et al.* 2014).

Soil respiration and T were exponentially correlated and Q_{10} (temperature sensitivity coefficient) generally decreased as soil depth increased. Q_{10} derived from changes in soil respiration and T may have increased or decreased during the measurement period over the different seasons, especially during winter and summer (Pingintha *et al.* 2010). Soil respiration generally increased in response to soil warming and decreased in the autumn when soil temperatures started to cool (Tucker and Reed 2016).

Both T and θ can influence CO₂ flux and the degree to which each affects soil respiration has been determined (Davidson *et al.* 1998, Law *et al.* 2001, Zimmermann *et al.* 2015). Soil CO₂ flux generally increases as water content increases but the effect on CO₂ is comprehensive through its influence on the respiratory activities of microbes and plant roots (Vargas and Allen 2008), and on the transport of gases through the soil (Li *et al.* 2016).

Unlike the CO₂ fluxes measured by the chambers, the dynamic model applied in our study could estimate the distribution of CO₂ fluxes and concentrations in the soil profile, suggesting that this model could provide an effective method for studying CO₂ flux and concentration on various temporal and spatial scales. A more comprehensive tool for studying the carbon cycle in forest ecosystems could be established by combining this model with a model describing tree growth and biomass allocation within the forest.

Soil CO₂ flux however, is dependent on CO₂ assimilation, the corresponding litter production, root exudates, and decomposition of soil organic matter and must be considered when calculating CO₂ flux by this model. Further research is needed to apply this model to different ecosystems. For example, the quantity and quality of soil organic matter and its seasonal distribution need to be better incorporated into the model to reflect seasonal variation of respiration (Moyes and Bowling 2013, Delogu *et al.* 2016).

Conclusions

This study provides a soil gradient method for calculating CO₂ flux by the continuous measurement of soil CO₂ concentrations using CO₂ sensors buried at different soil depths. The measurement of *in situ* CO₂ concentrations at a high temporal resolution provides information for the vertical distribution of CO₂ and its seasonal dynamics. T and soil respiration were exponentially correlated, indicating that the seasonal patterns of CO₂ flux were driven by temperature but water content may also be an important factor. Soil CO₂ flux determined from concentration profiles agree well with chambers, and suggest that the model to provide a practical and successful means of measuring soil CO₂ emissions and the important processes underlying soil CO₂ efflux maybe well apply in rocky mountain areas of northern China.

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