

## Soil CO<sub>2</sub> efflux of a larch forest in northern Japan

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**Abstract.** We had continuously measured soil CO<sub>2</sub> efflux ( $R_s$ ) in a larch forest in northern Japan at hourly intervals for the snow-free period in 2003 with an automated chamber system and partitioned  $R_s$  into heterotrophic respiration ( $R_h$ ) and autotrophic respiration ( $R_r$ ) by using the trench method. In addition, we applied the soil CO<sub>2</sub> concentration gradients method to continuously measure soil CO<sub>2</sub> profiles under snowpack in the snowy period and to partition  $R_s$  into topsoil (O<sub>a</sub> and A horizons) CO<sub>2</sub> efflux ( $F_t$ ) with a depth of 0.13 m and sub-soil (C horizon) CO<sub>2</sub> efflux ( $F_c$ ). We found that soil CO<sub>2</sub> effluxes were strongly affected by the seasonal variation of soil temperature but weakly correlated with soil moisture, probably because the volumetric soil moisture (30–40% at 95% confidence interval) was within a plateau region for root and microbial activities. The soil CO<sub>2</sub> effluxes changed seasonally in parallel with soil temperature in topsoil with the peak in late summer. On the other hand, the contribution of  $R_r$  to  $R_s$  was the largest at about 50% in early summer, when canopy photosynthesis and plant growth were more active. The temperature sensitivity ( $Q_{10}$ ) of  $R_r$  peaked in June. Under snowpack,  $R_s$  was stable until mid-March and then gradually increased with snow melting.  $R_s$  summed up to 79 gC m<sup>-2</sup> during the snowy season for 4 months. The annual  $R_s$  was determined at 934 gC m<sup>-2</sup> y<sup>-1</sup> in 2003, which accounted for 63% of ecosystem respiration. The annual contributions of  $R_h$  and  $R_r$  to  $R_s$  were 57% and 43%, respectively. Based on the gradient approach,  $R_s$  was partitioned vertically into litter (O<sub>i</sub> and O<sub>e</sub> horizons) with a depth of 0.01–0.02 m, topsoil and sub-soil respirations with proportions of 6, 72 and 22%, respectively, on an annual basis. The vertical distribution of CO<sub>2</sub> efflux was consistent with those of soil carbon and root biomass.

### 1 Introduction

The world's soils contain about 1550 Pg of organic carbon, which is more than twice the amount in the atmosphere (IPCC, 2007). Forests worldwide contain about 45% of the global carbon stock, a large part of which is in the forest soils. Recently, Bond-Lamberty and Thomson (2010) estimated that the global soil CO<sub>2</sub> efflux, widely referred to as soil respiration ( $R_s$ ), was about 98 Pg C y<sup>-1</sup> in 2008 based on a five-decade record of chamber measurements, which is more than 13 times the rate of fossil fuel combustion (IPCC, 2007), indicating that 20–40% of the atmospheric CO<sub>2</sub> circulates through soils every year. Overall,  $R_s$  is the largest component of ecosystem respiration (RE) and the second largest flux in the global carbon cycle after gross primary production (GPP).  $R_s$  is therefore a key process that is fundamental to our understanding of the terrestrial carbon cycle (Davidson and Janssens, 2006). A relatively small change in the carbon flow into or out of soils can strongly influence the global carbon cycle. For example, it was reported that the global  $R_s$  increased by 0.1 Pg C y<sup>-1</sup> between 1989 and 2008 (Bond-Lamberty and Thomson, 2010), and that the positive feedback from this enhancement of  $R_s$  by global warming would further raise atmospheric CO<sub>2</sub> concentration by 20–224 ppm by 2100 and resulting higher CO<sub>2</sub> levels would lead to an additional temperature increase ranging from 0.1 to 1.5 °C (Friedlingstein et al., 2006; IPCC, 2007).

In forest ecosystems, micrometeorological studies (i.e., eddy covariance) have shown that, on average, about 80% of GPP is respired back to the atmosphere (Law et al., 2002), and  $R_s$  has been estimated to account for 60–90% of RE, with marked temporal as well as spatial variations (Law et al., 1999; Janssens et al., 2001a, b; Liang et al., 2004). Therefore,  $R_s$  has recently received much attention from researchers and its accurate measurement is critical for developing a reliable model of carbon exchange in forest ecosystems (Jassal et al., 2007; Zhou et al., 2009).



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Because  $R_s$  has high spatial variability and the soil medium is not easily accessible,  $R_s$  cannot be measured by large-scale remote sensing. FLUXNET has become an effective network for observing carbon sequestration or loss by global terrestrial ecosystems by the eddy covariance technique (Luyssaert et al., 2008). Unfortunately, the use of the eddy covariance technique for measuring soil CO<sub>2</sub> efflux, especially below forest canopies, is often hampered by relatively low wind speeds (Drewitt et al., 2002) as well as by an abundance of understory vegetation (Lee, 1998; Janssens et al., 2001a). Therefore, to validate nocturnal, sub-canopy, and bad-weather (e.g., rainy period) eddy covariance measurements as well as the partition of the net ecosystem production (NEP), the flux research community have used automated chamber systems, which provide continuous (i.e., half-hourly or hourly) measurements of  $R_s$  (Gaumont-Guay et al., 2009; Jassal et al., 2007). Moreover, the automated continuous measurements of  $R_s$  (Goulden and Crill, 1997; Savage and Davidson, 2003; Liang et al., 2004) provide insights about ecosystem processes, which were not possible to explore before (Vargas et al., 2010).

Although larch forests are an important forest biome in Northeast Eurasian continent (Gower and Richards, 1990), continuous measurements of  $R_s$  are quite limited (Liang et al., 2004; Kim et al., 2005a). In this study, we continuously measured  $R_s$  at hourly intervals with an automated chamber system and partitioned it into heterotrophic and autotrophic soil respirations using the trench method in a larch forest in northern Japan. In addition, we applied the gradient method to measure  $R_s$  under snowpack in the winter season and to determine the vertical distribution of soil CO<sub>2</sub> efflux. The objectives of this study are to investigate (1) seasonal variations in soil CO<sub>2</sub> efflux, (2) responses of soil CO<sub>2</sub> efflux to temperature, moisture and rain events, (3) relative contributions of heterotrophic and autotrophic respirations to total  $R_s$ , (4) vertical distribution of soil CO<sub>2</sub> efflux and (5) contribution of  $R_s$  to RE, in a larch forest.

## 2 Site description

The study site is the Tomakomai Flux Site (42°44' N, 141°31' E), a larch plantation in Tomakomai National Forest, southern Hokkaido, Japan. The altitude of the site is 125 m and the terrain is essentially flat with a gentle slope of 1–2°. This site was one of the core sites of AsiaFlux network. The tower based eddy CO<sub>2</sub> flux had been measured using both open-path and closed-path systems since August 2000 until the catastrophic damage by a typhoon in September 2004 (Hirano et al., 2003a; Hirata et al., 2007).

### 2.1 Vegetation characteristics

The forest was a 45-year-old Japanese larch (*Larix kaempferi* Sarg.) plantation, interspersed with naturally generated Japanese spruce (*Picea jezoensis* Sieb. et Zucc.) and mixed broad-leaved species (*Betula* spp.). In 2001, for trees with breast height diameter (DBH) larger than 0.05 m, stand density was 1087 stems ha<sup>-1</sup> and basal area density was 23.2 m<sup>2</sup> ha<sup>-1</sup>, of which larch accounted for 81%. The forest canopy was about 15 m in height, and the overstory canopy leaf area index (LAI) reached its peak at 5.6 m<sup>2</sup> m<sup>-2</sup>. The forest floor was densely covered with perennial buckler fern (*Dryopteris crassirhizoma*) but lacked other understory species and moss. In late June, the average height, biomass, and LAI of the understory species were 0.5 m, 2.7 t ha<sup>-1</sup>, and 3.6 m<sup>2</sup> m<sup>-2</sup>, respectively. Defoliation started in mid-October, and the soil was covered by snow for about 4 months from mid-December to mid-April. Snow depth was 0.6 to 1.0 m in midwinter.

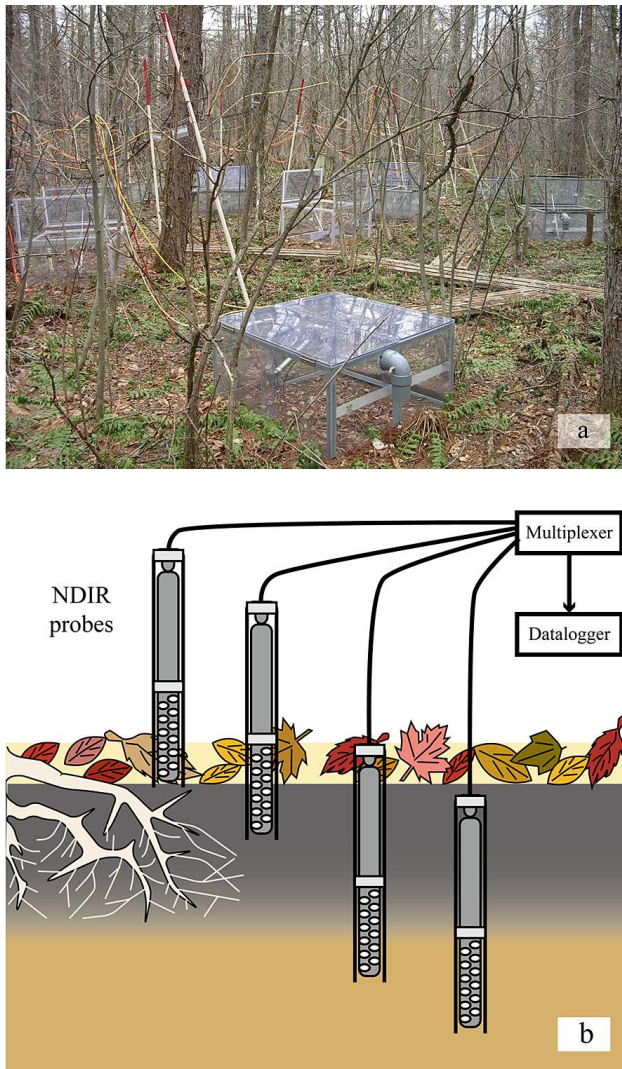
### 2.2 Climate

Climate records between 1979 and 2000 from two weather stations of Japan Meteorology Agency about 10 km away from the study site, Tomakomai and Shikotsuko, showed that the mean annual precipitation was approximately 1500 mm, and the mean annual temperature was 7.1 °C, with the mean monthly temperature ranging from -4.5 °C in January to 19.8 °C in August.

### 2.3 Soil characteristics

The soil is a homogeneous, well-drained, arenaceous soil developed from volcanoclastic sediment derived from a volcanic eruption that occurred about 300 years ago. It is classified as an immature Volcanogenous Regosol. The litter layer (O<sub>i</sub> and O<sub>e</sub> horizons) is 0.01–0.02 m thick and overlies a 0.1- to 0.15-m-thick organic layer (O<sub>a</sub> and A horizons) containing many fine roots. Beneath it, there is a layer composed of fragments of porous pumice stone (0.005–0.03 m in diameter) (C horizon) with some coarse roots. B horizon is lacking. The soil is weakly acidic (pH 5.0–6.0) and poor in nutrients. Sakai et al. (2007) reported that the densities of total and fine root biomass were 24.3 and 6.9 t ha<sup>-1</sup> and more than 80% of root biomass was distributed in the topsoil with a thickness of 0.15 m. Total soil organic carbon (SOC) and nitrogen storage were about 36 tC ha<sup>-2</sup> and 300 gN m<sup>-2</sup>, respectively, and about 90% of SOC accumulated in the surface layer between 0–0.30 m (Sakai et al., 2007).





**Fig. 2.** (a) Image of the multi-channel automated chamber systems installed at the site and (b) a set of infrared gas analyzers (IRGA sensors) vertically installed at different depths of soil.

where  $V$  is the effective chamber-head volume ( $\text{m}^3$ ),  $S$  is the measured soil surface area ( $\text{m}^2$ ),  $P$  is the air pressure (hPa),  $T$  is the air temperature (K), and  $W$  is the water vapor mole fraction ( $\text{mmol mol}^{-1}$ ) inside the chambers;  $\delta C/\delta t$  is the rate of change in the CO<sub>2</sub> mole fraction ( $\mu\text{mol mol}^{-1} \text{s}^{-1}$ ) calculated by the least-square method, and  $R$  is the gas constant ( $8.314 \text{ Pa m}^3 \text{ K}^{-1} \text{ mol}^{-1}$ ). Note that the pressure is not the pressure inside the IRGA cell but the pressure inside the chambers and we assumed that there was null pressure difference between the inside and outside of the chambers.

### 3.2 Soil CO<sub>2</sub> gradient system

We applied the soil CO<sub>2</sub> gradient method for continuous measurements of topsoil and subsoil CO<sub>2</sub> effluxes (Hirano et

al., 2003b; Liang et al., 2004; Tang et al., 2003). We inserted CO<sub>2</sub> probes (18.5 mm diameter, 155 mm long) of IRGAs (GMT222, Vaisala, Helsinki, Finland) into the soil to directly measure soil CO<sub>2</sub> concentration profiles. The probes were enclosed in polytetrafluoroethylene (PTFE) socks, which are porous water proofing (Liang et al., 2004), to prevent damage by rainwater, and then they were put into PVC (inner diameter 22 mm) casings to measure CO<sub>2</sub> concentrations at only specific soil depths. A fine (0.5 mm) mesh stainless steel screen was set at the opening at the bottom of the casing to prevent soil particles from entering the casing. In addition, temperature and pressure were measured in each casing with a thermocouple and a pressure transmitter (MPX4115, FreeScale Semiconductor, Austin, Texas, USA) via a plastic tube connected to the casing, respectively.

In June 2002, we installed two sets of sensors each with four CO<sub>2</sub> probes at two locations 0.60 m apart at 0, 0.02, 0.11 and 0.13 m depths. The measuring ranges of the probes were 0–2000 ppm for 0 and 0.02 m and 0–10000 ppm for 0.11 and 0.13 m. The probes were vertically installed to minimize soil disturbance (Fig. 2b). We defined 0 m depth as the soil surface under the litter layer ( $O_i$  and  $O_e$  horizons). To avoid heating of the soil adjacent to the probes, all probes were powered on hourly at 24 min past the hours, and operated for only 7 min. After warm-up of 5 min, CO<sub>2</sub> concentration was measured for 2 min and recorded every 10 s with a datalogger (CR10X). The temperature and pressure inside the casing were recorded simultaneously. Soil temperature at 0.02, 0.05, 0.10 and 0.15 m depths and volumetric soil moisture at 0.10 m depth were also recorded. The probes were removed for drying and calibration every two months.

The two replicates of CO<sub>2</sub> concentration were averaged at each depth and corrected for temperature and pressure accordingly to the manufacturer's instruction and converted to CO<sub>2</sub> molar density ( $\mu\text{mol m}^{-3}$ ). Then, soil CO<sub>2</sub> efflux ( $F_s$ ) was calculated by Fick's first law under the assumption of horizontal homogeneity,

$$F_s = -D_s \frac{\partial C}{\partial z} \quad (2)$$

where  $F$  is the CO<sub>2</sub> efflux ( $\mu\text{mol m}^{-2} \text{ s}^{-1}$ ),  $D_s$  ( $\text{m}^2 \text{ s}^{-1}$ ) is the gaseous CO<sub>2</sub> diffusion coefficient, and  $\delta C/\delta z$  is the vertical CO<sub>2</sub> density gradient ( $\mu\text{mol m}^{-4}$ ). Following Hirano et al. (2003b), we calculated soil CO<sub>2</sub> efflux from the soil surface ( $R_s$ ) and CO<sub>2</sub> efflux from C horizon to A horizon ( $R_c$ ) every hour from the soil CO<sub>2</sub> concentration profiles. We applied Eq. (2) to the boundary between  $O_e$  and  $O_a$  horizons (soil surface) and that between A and C horizons at a depth of 0.13 m, and considered CO<sub>2</sub> storage change in the  $O_a$  and A horizons for the calculation. In addition, we calculated CO<sub>2</sub> production rate of topsoil ( $O_a$  and A horizons) ( $R_t$ ) by subtracting CO<sub>2</sub> production rate of subsoil ( $R_c$ ) from  $R_s$ .  $D_s$

was calculated using Campbell's function (Campbell, 1985):

$$D_s = D_0 B \left( \frac{T_{\text{soil}} + 273.15}{273.15} \right)^{1.75} \left( \frac{1013}{P_s} \right), \quad (3)$$

where  $D_0$  is the CO<sub>2</sub> diffusion coefficient in air ( $1.39 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$ ) at 1013 hPa and 273.15 K,  $T_{\text{soil}}$  is the soil temperature (°C) at 0.02 or 0.10 m depth, and  $P_s$  is the air pressure (hPa) inside the sensor casing.  $B$  is the relative soil gaseous diffusion coefficient, which was determined from air-filled porosity ( $e$ ) of soil using their power relationship ( $B = 0.82e^{2.03}$ ,  $r^2 = 0.94$ ) (Currie, 1960). The relationship was obtained with undisturbed soil cores by the diffusion chamber method (Currie, 1960).  $e$  was determined as residual of the volume fractions of solid and water ( $\theta$ ). The  $D_s$  in A horizon fluctuated between 0.015 and 0.045  $\text{cm}^2 \text{ s}^{-1}$  with soil temperature and moisture, which is in the middle range of 0.022 to 0.052  $\text{cm}^2 \text{ s}^{-1}$  for Japanese forest soils (42 ecosystems) (Ishitsuka and Sakata, 2006).

### 3.3 Data analysis

Hourly measurements of the 8 chambers used for each process ( $R_s$  and  $R_h$ ) were averaged to obtain the mean efflux, and coefficients of variation (CV) were calculated to show spatial variation. To examine the temperature response of soil CO<sub>2</sub> efflux, we performed a regression analysis using an exponential model:

$$R_s = ae^{bT_{\text{soil}}}, \quad (4)$$

where  $R_s$  is the CO<sub>2</sub> efflux at soil temperature  $T_{\text{soil}}$  at a depth of 0.05 m,  $a$  is the efflux at 0°C, and  $b$  is the sensitivity of the soil CO<sub>2</sub> efflux to temperature. The value of  $b$  was also used to calculate the  $Q_{10}$  coefficient:

$$Q_{10} = e^{10b}, \quad (5)$$

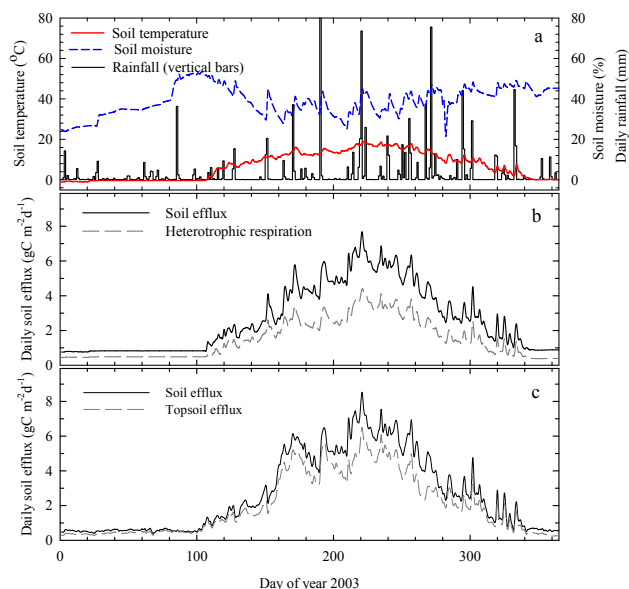
which is the relative increase in CO<sub>2</sub> efflux with a 10°C increase in soil temperature.

Although chamber data were missing for the snow-covered period for 127 days from January 1 to April 17 and December 12 to December 31 in 2003, more than 95% of data were available for the snow-free period. Also, more than 95% of gradient flux data were available throughout the year. The data gaps were filled with  $R_s$  estimated from Eq. (4) by using soil temperature. Root respiration ( $R_r$ ) was estimated by subtracting  $R_h$  from  $R_s$ .

## 4 Results and discussion

### 4.1 Seasonal variations in soil CO<sub>2</sub> efflux ( $R_s$ )

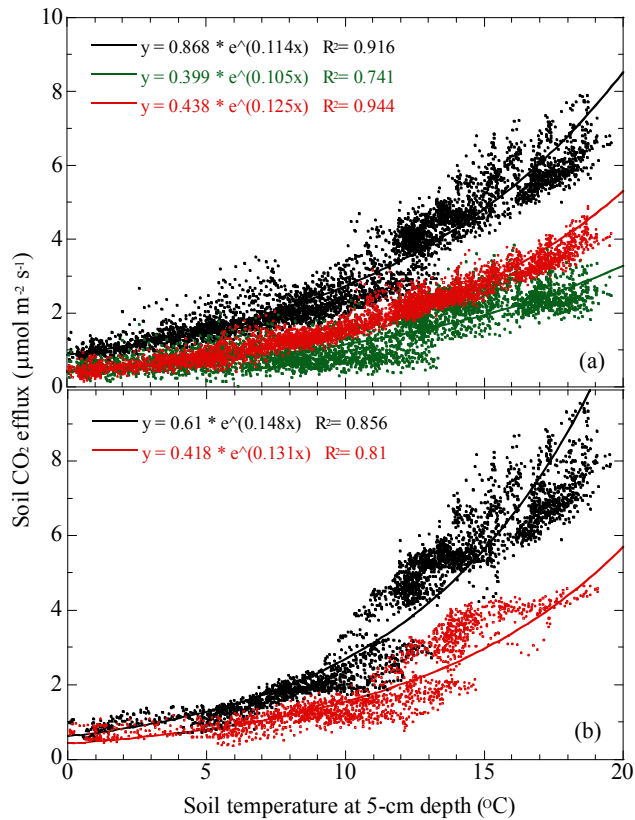
$R_s$  measured by both the chamber and gradient systems showed notable seasonal patterns (Fig. 3). After snow disappearance on 17 April  $R_s$  increased rapidly until around DOY



**Fig. 3.** Seasonal variations in daily means or sums of (a) soil temperature at a depth of 0.05 m (solid line), precipitation (bars) and volumetric soil moisture at a depth of 0.1 m (dashed line), (b) total soil respiration (solid line) and heterotrophic respiration (gray dashed line) measured with the automated chamber system, (c) and total soil respiration (solid line) and topsoil (O<sub>a</sub> and A horizons) efflux (gray dashed line) measured with the gradient system. For the chamber method, measurements were conducted between DOY 108 and 345. Other values were estimated by using the  $Q_{10}$  function of Eq. (4).

(day of year) 220 with soil temperature.  $R_s$  remained high during late summer between DOY 220 and 260. Peak  $R_s$  values were about 6 and 7  $\text{gC m}^{-2} \text{ d}^{-1}$  by the chamber and the gradient systems, respectively. Then,  $R_s$  decreased steadily with soil temperature until the ground was covered by snow. The pattern in the snow-free season is consistent with that in 2001 (Liang et al., 2004) and RE patterns with the peak in August (Hirata et al., 2007). Heterotrophic respiration ( $R_h$ ) showed a similar seasonal variation to that of  $R_s$ . Also, topsoil respiration ( $F_t$ ) varied almost in parallel with  $R_s$ .

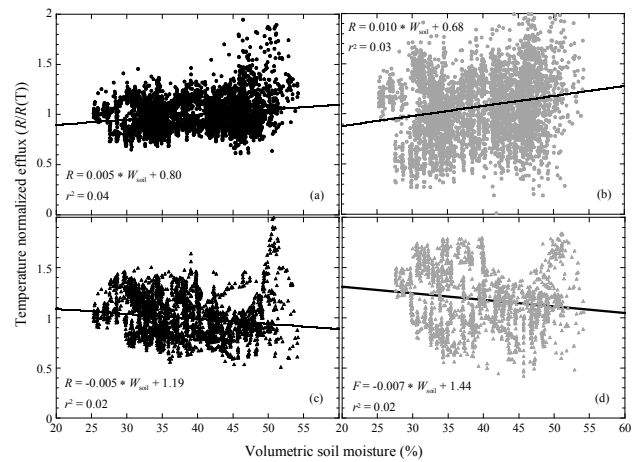
During the snowy season,  $R_s$  measured by the gradient method was stable at 0.50–0.55  $\mu\text{mol m}^{-2} \text{ s}^{-1}$  until mid-March and then gradually increased up to 0.8  $\mu\text{mol m}^{-2} \text{ s}^{-1}$  with snow melting. Mean  $R_s$  for the snowy season of 127 days was  $0.56 \pm 0.12 \mu\text{mol m}^{-2} \text{ s}^{-1}$  (mean  $\pm$  SD). The seasonal variation and the mean value are very similar to those measured in a neighboring deciduous broadleaf forest using the gradient method (Hirano, 2005). On the other hand,  $R_s$  can be also estimated from soil temperature using Eq. (4) fitted to chamber data. The mean and SD of the estimated value was  $0.83 \pm 0.03 \mu\text{mol m}^{-2} \text{ s}^{-1}$ . However, compared to the patterns of the gradient method, there was no variation during the snow-melting season. This stable  $R_s$  is due to the stable soil temperature under snowpack.



**Fig. 4.** Relationships between soil temperature at a depth of 0.05 m and (a) total soil CO<sub>2</sub> efflux (black symbols and line), heterotrophic (red symbols and line) and root (green symbols and line) respirations measured by the chamber method, and (b) total soil CO<sub>2</sub> efflux (black symbols and line) and topsoil (O<sub>a</sub> and A horizons) CO<sub>2</sub> efflux (red symbols and line) measured by the gradient technique, on an hourly basis. The best-fitted exponential curves are shown.

#### 4.2 Responses of soil CO<sub>2</sub> efflux to soil temperature and moisture

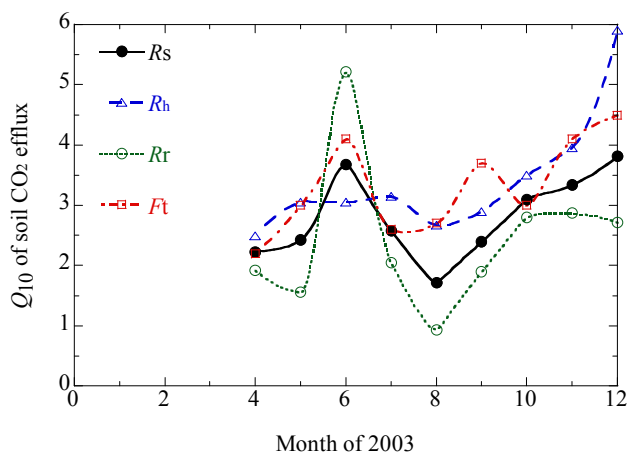
For the snow-free period for 8 month (Fig. 4), soil CO<sub>2</sub> effluxes increased exponentially with temperature and  $Q_{10}$  value was estimated to be about 3.1, 3.5, 2.9 and 3.6 for  $R_s$ ,  $R_h$ ,  $R_r$  and  $F_t$ , respectively. The  $Q_{10}$  of  $R_s$  was quite larger than the global mean  $Q_{10}$  of  $R_s$  (ranging 1.43–2.03; Raich et al., 2002; Zhou et al., 2009). Generally,  $Q_{10}$  varies between 1 and 5, and is negatively correlated with temperature and positively correlated with soil moisture (Lloyd and Taylor, 1994). Therefore, the high  $Q_{10}$  is attributable mainly to high soil moisture, ranging between 25 and 55% with 95% confidence interval of 30–40%, and relatively low air temperature with an annual mean of 7.1°C. The  $Q_{10}$  was lower for  $R_r$  than for  $R_h$  in this site. The result is inconsistent with the report for a temperate mixed forest at the Harvard Forest, in which the  $Q_{10}$  of  $R_r$  (4.6) was significantly greater than that of  $R_h$  (2.5) (Boone et al., 1998). Our finding suggests that



**Fig. 5.** Temperature normalized soil CO<sub>2</sub> efflux, ratio between measured soil CO<sub>2</sub> efflux ( $R_s$ ) and its temperature fitted value ( $R_s(T)$ ), versus volumetric soil water content at a depth of 0.1 m. (a) and (b) represent total soil CO<sub>2</sub> efflux and root respiration measured by the automated chamber system, and (c) and (d) represents total soil CO<sub>2</sub> efflux and topsoil (O<sub>a</sub> and A horizons) CO<sub>2</sub> efflux measured by the soil CO<sub>2</sub> gradient system.

the high temperature sensitivity of  $R_h$  will offset the forest carbon sequestration in the changing world under elevated atmospheric CO<sub>2</sub> concentration (Bond-Lamberty and Thomson, 2010). The  $Q_{10}$  of  $F_t$  from O<sub>a</sub> and A horizons with a depth of 0.13 m was almost the same as that of  $R_h$ , whereas it was larger than that of  $R_s$ .  $R_s$  includes CO<sub>2</sub> efflux from litter layer (O<sub>e</sub> and O<sub>i</sub> layers) with low  $Q_{10}$  values around 2 (Kim et al., 2005b) and that from C horizon, which is almost independent of temperature at a depth of 0.5 m (Hirano et al., 2003b). This fact can explain the difference in  $Q_{10}$  between  $R_s$  and  $F_t$ .

Soil CO<sub>2</sub> efflux is also controlled by moisture availability. To remove the confounding effect of temperature and evaluate the role of soil moisture on soil CO<sub>2</sub> efflux, we plotted temperature-normalized efflux (i.e., the ratio of observed soil CO<sub>2</sub> efflux to temperature-fitted efflux) against volumetric soil moisture (Fig. 5). Temperature-normalized values were largely scattered around 1.0 throughout the snow-free period. Low coefficients of determination ( $R^2 < 0.04$ ) and slight slopes ( $\leq 0.01$ ) of regression lines suggest that, soil moisture of 30–40% at 95% confidence interval favored root and microbial activities. For this forest, there was indeed no clear correlation between soil moisture and GPP or RE that was observed based on the flux tower measurement (Hirata et al., 2007). However, the larger deviations at high soil moisture were probably induced by the rain events (Fig. 5). Results are consistent with the findings in relative humid forest ecosystems (Tang et al., 2008; Ruehr et al., 2010; Klimek et al., 2009).

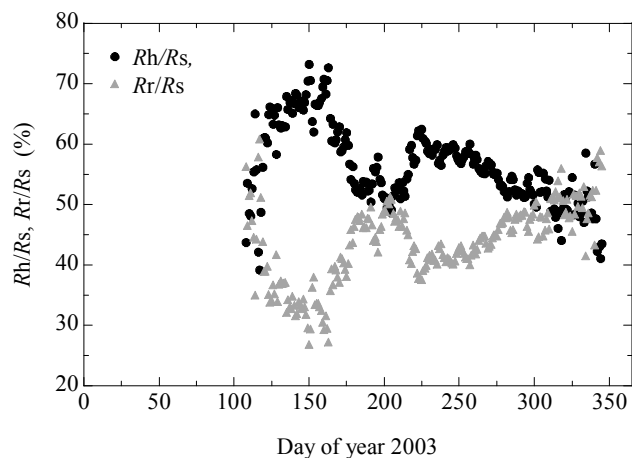


**Fig. 6.** Seasonal changes in  $Q_{10}$  of soil CO<sub>2</sub> efflux ( $R_s$ , dots with solid line), heterotrophic respiration ( $R_h$ , triangles with dashed line), root respiration ( $R_r$ , circles with dotted line), and topsoil CO<sub>2</sub> efflux ( $F_t$ , squares with dot-dashed line) for the snow-free season from April through December.

#### 4.3 Seasonal variations in $Q_{10}$ of soil CO<sub>2</sub> efflux

To investigate the mechanism by which temperature affects soil CO<sub>2</sub> efflux, we determined  $Q_{10}$  values monthly. Figure 6 shows seasonal variations in  $Q_{10}$  of  $R_s$ ,  $R_h$ ,  $R_r$  and  $F_t$ . The  $Q_{10}$  of  $R_r$  peaked in June when productivities of fine roots and the rhizosphere were highest, suggesting that  $R_r$  is controlled mainly by canopy processes (e.g. photosynthesis) through metabolism of recently fixed carbohydrates (Tang et al., 2005; Moyano et al., 2008; Sampson et al., 2007; Irvine et al., 2008; Baldocchi et al., 2006; Yuste et al., 2004; Haselquist et al., 2010). The fact that ecosystem photosynthesis (GPP) of this larch forest also peaked in June (Hirata et al., 2007) supports this suggestion.  $R_r$  showed the lowest  $Q_{10}$  in August (i.e., midsummer). In contrast, the  $Q_{10}$  of  $R_h$  remained relatively constant at around 3 throughout the growing season but increased drastically from late autumn (October) to early winter (December), accompanied by a large decrease in temperature. The increase in  $Q_{10}$  of  $R_h$  in late autumn was attributable to fresh litter supply through defoliation (Kim et al., 2005a). Precipitation was abundant for the growing season in this site, and no evidence for seasonal drought was observed. Thus, the different seasonality of  $Q_{10}$  between  $R_r$  and  $R_h$  suggests that the temperature response of  $R_r$  has a different mechanism from that of  $R_h$ .

There is an increasing evidence that  $Q_{10}$  of  $R_s$  is not seasonally constant and tends to increase with decreasing temperature and increasing soil moisture (Chen et al., 2009a). Recent field studies have also observed significant seasonal variations in  $Q_{10}$  of  $R_s$  (Chen et al., 2009a; Janssens and Pilegaard, 2003; Liu et al., 2006b; Phillips et al., 2010; Xu and Qi, 2001). By partitioning  $R_s$  into  $R_r$  and  $R_h$ , we found that the seasonality of the  $Q_{10}$  of  $R_s$  was almost parallel with that



**Fig. 7.** Seasonal variations in the contributions of heterotrophic respiration ( $R_h$ , triangles) and root respiration ( $R_r$ , circles) to total soil respiration ( $R_s$ ) measured by the chamber method on daily basis.

of the  $Q_{10}$  of  $R_r$  (Fig. 6). The summer depression in  $Q_{10}$  was also reported by Janssens and Pilegaard (2003), whereas they suspected that the depression was caused by summer drought stress. The similar seasonal variations in the  $Q_{10}$  of  $R_s$  and  $R_r$  suggest that the large seasonal change in  $R_r$  dominates the seasonal pattern of  $R_s$  during the growing season. The relative stability of the  $Q_{10}$  of  $R_h$  is consistent with both laboratory results and theoretical predictions (Davidson and Janssens, 2006). The seasonal  $Q_{10}$  variation indicates that a  $Q_{10}$  function based on annual data will under- or overestimate  $R_s$  on shorter timescales. Thus, empirical models should be parameterized at a time resolution similar to that required by the output of each model.

#### 4.4 Root respiration ( $R_r$ ) and heterotrophic respiration ( $R_h$ )

Distinguishing  $R_r$  from  $R_h$  is an important first step in interpreting field measurements, because  $R_r$  and  $R_h$  can respond differently to the environment (Cisneros-Dozal et al., 2007; Moyano et al., 2008; Irvine et al., 2008; Ryan and Law, 2005). In this larch forest, the relative contributions of  $R_r$  and  $R_h$  to  $R_s$  showed distinct seasonal patterns (Fig. 7), which is consistent with the suggestion of Hanson et al. (2000) that the proportions of  $R_r$  and  $R_h$  to  $R_s$  vary seasonally and among ecosystems.

$R_h$  accounted for most of  $R_s$  (65–70%) between late April and early June (DOY 120 to 160), probably because a rapid increase in soil temperature after the thaw enhanced decomposition of the leaf litter accumulated in the last autumn, as well as because the decreasing soil moisture led to increase soil oxygenation, which stimulated microbial activity. As a result, the contribution of  $R_h$  to  $R_s$  increased, and that of  $R_r$  decreased in this period. After the canopy began to leaf out, from early May, GPP rapidly increased and was maintained

at a high rate from early June through mid-August (Hirata et al., 2007), which probably provided substrate for root and associated rhizosphere (Vargas et al., 2010; Tang et al., 2005; Moyano et al., 2008; Hasselquist et al., 2010). Thus,  $R_r$  contributed more (around 50%) to  $R_s$  between early June and early August (DOY 161 to 215), when the plants grew rapidly. From early August to mid-September (DOY 216 to 258), high temperatures probably both inhibited photosynthesis and enhanced decomposition of organic matter, allowing the heterotrophic contribution to reach a second peak. As the temperature decreased from the beginning of September, the  $R_h$  contribution decreased but the  $R_r$  contribution remained at a relatively high level owing to the higher allocation of photosynthate (i.e., starch) to roots and ectomycorrhizal fungi (Liang et al., 2004; Liu et al., 2006a; Kurganova et al., 2007; Hasselquist et al., 2010). In the end of September, the contributions of  $R_r$  and  $R_h$  became equal, which lasted until December. On the other hand, the contribution of  $F_i$  to  $R_s$  showed a simpler seasonal variation (data are not shown). The  $F_i$  contribution linearly increased from 60% to 90% by late April for 4 months and gradually decreased to 70% by late November. During December, it decreased to 45%. The seasonal variation is similar with that in a neighboring deciduous broadleaf forest (Hirano et al., 2003b; Hirano, 2005).

During the whole growing season for 5 months, between DOY 135 and 288, the average contribution of  $R_h$  to  $R_s$  was 57%. During the non-growing season,  $R_r$  and  $R_h$  each accounted for roughly half of  $R_s$ . Hanson *et al.* (2000) reviewed that, in forests, heterotrophic contributions were ranging from 40% during the growing season to 54% annually. The root exclusion method by trenching, which was used in this study to distinguish  $R_r$  from  $R_h$ , can overestimate  $R_h$  in the short term (e.g., within the first treatment year) owing to the decomposition of dead roots, whereas it can underestimate  $R_h$  in the long term (e.g., over one year) because no new fine root litter is supplied.

#### 4.5 Impact of rainfall events on soil CO<sub>2</sub> efflux

Several studies have detected a sudden increase in soil respiration during pulsed rain events especially in arid ecosystems (Lee et al., 2004; Xu et al., 2004; Irvine et al., 2005; Kelliher et al., 2004; Chen et al., 2009b; Inglema et al., 2009; Baldocchi et al., 2006). In this study, both the automated chamber and the gradient approaches revealed episodic CO<sub>2</sub> emissions (Fig. 3).  $R_s$  rapidly responded to the onset of rain and increased by approximately 70% following a rain event with more than 20 mm of precipitation. After the rain,  $R_s$  returned to the pre-rain rate in several hours. Our results are consistent with the findings of rain simulation studies (Lee et al., 2004; Chen et al., 2009b), and field observations (Inglema et al., 2009; Xu et al., 2004; Baldocchi et al., 2006). Lee et al. (2004) reported that a 170-mm rainstorm enhanced  $R_s$  up to fivefold, and  $R_s$  returns to the pre-rain level within

one hour after the rain. Kelliher et al. (2004) reported that, in a young ponderosa pine forest,  $R_s$  was increased threefold by a simulated rain event and returned to the pre-rain levels within 24 h after the rain. In the same forest, Irvine and Law (2002) showed that the intensity of rain events had a substantial effect on interannual variation in  $R_s$ , because heavy rain events resulted in prolonged elevation of  $R_s$ .

The rapid response of  $R_s$  to pulsed rain events suggests that continuous measurements are crucial for accurate, quantitative assessment of  $R_s$  (Vargas et al., 2010). Periodic measurements by the manual chamber method, which are generally conducted only under fine-weather conditions, undoubtedly underestimate  $R_s$  under rainy conditions, because the pulse signals of  $R_s$  are missed. The underestimation can strongly affect ecosystem carbon balance (Lee et al., 2004; Xu et al., 2004; McCulley et al., 2007). For instance, if we estimate annual  $R_s$  from soil temperature using two exponential models (Eq. 4) fitted to data on all days and no-rain days, respectively, the latter model ignoring rainy data underestimates annual  $R_s$  by 24 gC m<sup>-2</sup> or 2%, which is equivalent to 11% of the annual NEE of this forest (Hirata et al., 2007).

#### 4.6 Seasonal and annual sums of soil CO<sub>2</sub> efflux ( $R_s$ )

$R_s$  measured with the chamber system summed up to 855 gC m<sup>-2</sup> for about 8 months of the snow-free season. The seasonal sum of  $R_s$  measured by the gradient method for the same period was 896 gC m<sup>-2</sup>, whereas it must be underestimated, because the gradient method could not measure CO<sub>2</sub> efflux from the litter layer ( $Q_i$  and  $Q_e$  horizons). CO<sub>2</sub> efflux from the litter layer at this site can be estimated at 60 gC m<sup>-2</sup> for the snow-free season in 2002 from the results of Kim et al. (2005a). If its interannual variation is negligible,  $R_s$  comes to 956 gC m<sup>-2</sup> by adding the litter contribution, which is larger than that by the chamber method by 12%. The overestimation by the gradient method is consistent with the result in 2001 (Liang et al., 2004). However, it is worth noting that the difference between the two methods is considerably smaller than those reported by, for example, Vargas et al. (2008) and Pingintha et al. (2010), who showed that the gradient effluxes were 23% and 90% larger than that by the chamber method, respectively. In contrast, Baldocchi et al. (2006) reported that gradient effluxes were only 77% of chamber measurements. Because each hourly  $R_s$  by the gradient method always fell within the range of 8 measurements of  $R_s$  by the chamber method, the difference between the two methods is attributable to spatial variation in  $R_s$ ; the coefficient of variation (CV) was 21% for  $R_s$  and 20% for  $R_h$  on average.

For 4 months of the snowy season, the gradient-based  $R_s$  summed up to 73 gC m<sup>-2</sup>, which is almost the same as that in a neighboring deciduous broadleaf forest (Hirano, 2005). By adding litter CO<sub>2</sub> efflux (Kim et al., 2005a),  $R_s$  under snowpack is estimated to be 89 gC m<sup>-2</sup>. On the other hand, the chamber-based  $R_s$  may be used to estimate winter  $R_s$

by extrapolating Eq. (4) fitted for the snow-free season only, which overestimated the winter  $R_s$  by 26 gC m<sup>-2</sup>. As a result, when the chamber method and the gradient method are adopted for the snow-free and snowy seasons, respectively, the annual  $R_s$  is determined to be 934 gC m<sup>-2</sup> y<sup>-1</sup>. The annual  $R_s$  stands in the middle class of that of temperate forest (Bond-Lamberty and Thomson, 2010). The winter  $R_s$  for 4 months accounted for 8.5% of the annual  $R_s$ . In a neighboring broadleaf forest, the winter  $R_s$  accounted for 8–10% of the annual  $R_s$  (Hirano 2005; Kim et al., 2005a). The contribution of the winter  $R_s$  was slightly larger than those of forests, shrubs and meadow grasslands in North China (Wang et al., 2010). In our earlier study (Liang et al., 2004), we obtained an annual  $R_s$  of 665 gC m<sup>-2</sup> y<sup>-1</sup> with a steady-state chamber system at this site in 2001. The large difference of 269 gC m<sup>-2</sup> in the annual  $R_s$  may be attributed to the positive pressure of 0.22 Pa inside the steady-state chamber system, which led to a systematic underestimation of  $R_s$  (Fang and Moncrieff, 1998; Widen and Lindroth, 2003).

Based on the gradient measurement, the relative contribution to the annual  $R_s$  from topsoil, subsoil, and litter was 72, 22 and 6%, respectively. Such a proportion indicates that the organic soil layer with a depth of 0.13 m produced 72% of total CO<sub>2</sub> emitted to the atmosphere. The vertical distribution of CO<sub>2</sub> efflux can be explained by those of soil carbon content and root biomass, which were localized in surface soil (Sakai et al., 2007). The annual ecosystem respiration (RE) of this site was 1493 gC m<sup>-2</sup> y<sup>-1</sup> in 2003 (Hirata et al., 2007). The annual  $R_s$  of 934 gC m<sup>-2</sup> y<sup>-1</sup> accounted for 63% of the RE. The ratio of  $R_s$  to RE is compatible with those of European forests (Janssens et al., 2001b). Liang et al. (2005) reported that the annual sum of the stem respiration of larch trees was 78–80 gC m<sup>-2</sup> y<sup>-1</sup> in this forest in 2002. Considering the contribution of broadleaf species, the total stem respiration may be estimated to be about 100 gC m<sup>-2</sup> y<sup>-1</sup>. Then, we can attribute the residual of RE (i.e., 459 gC m<sup>-2</sup> y<sup>-1</sup>) to the aboveground respiration of tree canopy and understory species as well as the decomposition of coarse woody debris.

## 5 Conclusions

The concurrent employment of gradient method along with chamber method provide not only the soil effluxes during the snow-covered period but also further insight into their partitioning. Soil CO<sub>2</sub> effluxes of the larch forest, including total soil respiration ( $R_s$ ), heterotrophic respiration ( $R_h$ ), autotrophic (root) respiration ( $R_r$ ) and top soil respiration ( $F_t$ ), were strongly affected by soil temperature. Overall, no constraints of soil moisture on any soil CO<sub>2</sub> efflux were found, whereas soil respiration showed pulsed increase after rain events. The contribution of  $R_r$  to  $R_s$  changed seasonally, and peaked in early summer when canopy photosynthesis and plant growth were active. The vertical distribution of soil CO<sub>2</sub> efflux reflected those of soil carbon content and

root biomass. On an annual basis,  $R_r$  accounted for 43% of  $R_s$ , and  $R_s$  accounted for 63% of ecosystem respiration measured by the eddy covariance technique.

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