Carbon dioxide exchange between the atmosphere and an alpine meadow ecosystem on the Qinghai–Tibetan Plateau, China

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Abstract

We used the eddy covariance method to measure the CO2 exchange between the atmosphere and an alpine meadow ecosystem (37°29′–45°45′N, 101°12′–23′E, 3250 m a.s.l.) on the Qinghai–Tibetan Plateau, China in the 2001 and 2002 growing seasons. The maximum rates of CO2 uptake and release derived from the diurnal course of CO2 flux (F CO2 ) were −10.8 and 4.4 μmol m⁻² s⁻¹, respectively, indicating a relatively high net carbon sequestration potential as compared to subalpine coniferous forest at similar elevation and latitude. The largest daily CO2 uptake was 3.9 g C m⁻² per day on 7 July 2002, which is less than half of those reported for lowland grassland and forest at similar latitudes. The daily CO2 uptake during the measurement period indicated that the alpine ecosystem might behave as a sink of atmospheric CO2 during the growing season if the carbon lost due to grazing is not significant. The daytime CO2 uptake was linearly correlated with the daily photosynthetic photon flux density each month. The nighttime averaged F CO2 showed a positive exponential correlation with the soil temperature, but apparently negative correlation with the soil water content.

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1. Introduction

Carbon cycle studies in terrestrial ecosystems have helped identify the magnitude of sinks and sources of atmospheric CO2 (Wofsy et al., 1993; Oechel et al., 1993; Gesselen et al., 1996). The eddy covariance method is one of the most useful methods to measure the carbon budget between the atmosphere and various ecosystems (Falge et al., 2002a,b).

On a global basis, grasslands are one of the most widespread vegetation types. Natural grasslands, including tundra, cover about 30% of the Earth’s surface and contain 452.3 Pg (1/4 of the Earth’s total amount) of organic carbon in both biomass and soil (Adams et al., 1990). Scurlock and Hall (1998) reported that temperate and tropical grasslands sequester 0.5 Pg of carbon a year, and suggested that grassland ecosystems...
might play an important role as a sink of atmospheric carbon. However, most studies of the carbon budget have been conducted only on grassland ecosystems at elevations below 1500 m. There is little information on carbon flux for alpine grassland ecosystems at high elevations.

In alpine meadow ecosystems on the Qinghai-Tibetan Plateau in the western part of China, abundant light and precipitation allow plants to grow efficiently (i.e. sequester carbon) during the short growing season from May to September. The low temperature in winter may restrict the decomposition of litter and ecosystem respiration. Therefore, the alpine ecosystem may act as an annual net sink of atmospheric CO2 if the carbon lost from winter ecosystem respiration and grazing impacts is relatively small.

Alpine meadow ecosystems in China cover approximately 6.37 × 10^5 km^2 and contain 11.3 Pg of carbon in biomass and soil (Ni, 2002). The soil carbon density of this ecosystem (18.2 kg m^−2; Ni, 2002) is much higher than that of savanna (5.4 kg m^−2; Adams et al., 1990) and temperate grassland (13.0 kg m^−2; Adams et al., 1990) where low temperature limits the growing season. However, tundra is reported to be at risk of changing from a sink to a source of atmospheric CO2 with global warming (Oechel et al., 1993). Studies of alpine meadow ecosystems will further our understanding of carbon dynamics and help predict their responses to climate change.

In this study, we focused on the CO2 exchange during the growing season between the atmosphere and an alpine meadow ecosystem on the Qinghai-Tibetan Plateau. We determined the diurnal and seasonal changes in CO2 flux and their dependence on environmental factors, especially soil water content and soil temperature.

2. Sites and measurement

2.1. Site description

Measurements of CO2 flux (F CO2 ) and environmental factors were conducted at the Haibei Alpine Meadow Ecosystem Research Station, Northwest Plateau Institute of Biology, Chinese Academy of Science (37°29′-45°N, 101°12′-23°E; 3250 m a.s.l.). The research station is located in northeast Tibet in a large valley oriented northwest–southeast and surrounded by the Qilian Mountains. The average altitude of the mountains is 4000 m and of the valley is 2900–3500 m. The Datong River passes to the south. The landscape is characterized by large mountain ranges with steep valleys and gorges interspersed with relatively level and wide inter-mountain grassland basins.

The climate at Haibei Station is characterized by low temperature and limited precipitation. The annual average temperature and precipitation for 1981–2000 were -1.7 °C and 561 mm. In the growing season of May through September, the plentiful sunshine and rainfall (80% of annual total rainfall) allow plants to grow efficiently, although harmful UV-B radiation is high (Li and Zhou, 1998).

The soil is a clay loam; its average thickness is 0.65 m. The surface 0.05-0.10 m horizons, which are classified as Mat Cry-gelic Cambisols according to the Chinese national soil survey classification system (The Institute of Soil Science and the Chinese Academy of Sciences, 2001), are wet and high in organic matter.

The plant community is dominated mainly by three perennial sedges, Kobresia humilis, K. pygmaea, and K. tibetica (Cyperaceae), and by one dwarf shrub species, Potentilla fruticosa (Rosaceae) (Li and Zhou, 1998). The plants start to grow in May, when the air temperature starts rising, and reach maximum aboveground biomass (of 342 g d.w. m^−2; average for 1980–1993) in July and August, when the air temperature and precipitation are the highest of the year. Their aboveground parts die in October. During the growing season, the plants accumulate photosynthates in belowground storage organs in preparation for the next spring (Li and Zhou, 1998).

The study site is grazed by yaks and sheep every winter. Measurements were conducted in summer only, so we did not account for the effect of livestock respiration or grazing on plant growth.

2.2. Eddy covariance and microclimate measurements

CO2 and H2O fluxes were measured by the eddy covariance method from 9 August to 30 September 2001 and from 23 May to 30 September 2002. The study site is flat with a fetch of at least 250 m in all directions. Wind speed and sonic virtual temperature
were measured at 2.2 m above the ground with a sonic anemometer (CSAT-3, Campbell Scientific Inc., Logan, UT, USA). Carbon dioxide and water vapor concentrations were also measured at the same height with an open-path infrared gas analyzer (CS-7500, Campbell Scientific Inc.).

Other micrometeorological measurements were conducted at the same site. Net radiation and photosynthetic photon flux density (PPFD) were measured at 1.5 m above the ground with a net radiometer (CNR-1, Kipp & Zonen Inc., Saskatoon, Saskatchewan, Canada) and a quantum sensor (LI-190SB, Li-Cor, Lincoln, NE, USA). Air temperature and humidity were measured at 1.1 and 2.2 m above the ground with a humidity and temperature probe (HMP45C, Vaisala, Helsinki, Finland). Wind speed was measured at 1.1 and 2.2 m above the ground with cup anemometers (034A-L and 014A, R.M. Young Co., Traverse, MI, USA). Soil heat flux was measured at 0.02 m below the ground at three locations with heat flux plates (HFT-3, Campbell Scientific Inc.). Soil temperature was measured at 0.025, 0.05, 0.1, 0.2, 0.3, 0.4, 0.5, 0.6, and 0.7 m below the ground with copper–constantan thermocouples. Soil water content (SWC) was measured at 0.05, 0.2, and 0.5 m with time-domain reflectometry sensors (CS-615, Campbell Scientific Inc.). Soil surface temperature was measured at three points in a 1-m² area with thermistor thermometers (107 probe, Campbell Scientific Inc.). Rainfall was measured at 0.7 m above the ground with a tipping bucket (TE525MM, Campbell Scientific Inc.).

Fifteen-minute averages of all data were logged by an analog multiplexer (AM416) and a digital micrologger (CR23X, Campbell Scientific Inc.). Wind speed, sonic virtual temperature, and CO₂ and H₂O concentrations were sampled by the digital micrologger at 10 Hz. Examination of cospectra between vertical wind \( w' \) and CO₂ concentration \( c' \) indicated that the flux loss was insignificant with 15 min averaging. The mean, variance, and covariance values were then calculated and logged for every 15 min. The WPL density correction was then applied to fluxes of CO₂ and water vapor (Webb et al., 1980). Exponential regression of nighttime \( F_{CO_2} \) with \( u_* > 0.2 \text{ m s}^{-1} \) against soil temperature at a depth of 0.05 m every 2 weeks was used to fill data gaps for nighttime. The daytime missing data were extrapolated from a fortnightly regression against PPFD and soil temperature 0.05 m below the ground (Suyker and Verma, 2001).

2.3. Aboveground biomass and LAI measurements

Aboveground biomass and LAI were investigated within a radius of 250 m around the measuring station once in 2001 (20 August) and nine times during May–September in 2002. The aboveground biomass was cut at ground levels in five randomly placed 0.5 m² quadrates. LAI was then measured with an LAI meter (LI-3100, Li-Cor). After oven drying for 48 h at 70 °C, the dry matter was weighed on an electric balance.

2.4. Calculation of equilibrium ET and canopy resistance

Equilibrium evaporation is defined as the evaporation rate attained by a free surface after it saturates the atmosphere (Penman, 1948). Although this condition is rarely met in the field, comparing measured evaporation with predicted equilibrium evaporation helps diagnose the balance between supply capacity and atmospheric demand for evaporation. Equilibrium evaporation is defined as follows (e.g., Jones, 1992):

\[
\text{I}_{\text{eq}} = \frac{\Delta(R_n - G)}{\Delta + \gamma},
\]

where \( \text{I}_{\text{eq}} \) is the equilibrium evaporation (in W m⁻²), \( \Delta \) the slope of the relation between saturation vapor pressure and temperature (in hPa °C⁻¹), \( R_n \) the net radiation (in W m⁻²), \( G \) the soil heat flux (in W m⁻²), and \( \gamma \) the psychrometric constant (in hPa °C⁻¹).

Canopy resistance, \( R_{\text{canopy}} \) (i.e., the stomatal resistance of a “big leaf”), is determined by the physiological activity of the canopy plants in the Penman–Monteith equation (Monteith, 1972). By rearranging the equation, \( R_{\text{canopy}} \) (in s m⁻¹) is derived from environmental factors as follows:

\[
R_{\text{canopy}} = R_{\text{air}} \left[ \frac{\Delta(R_n - G - \text{I}_{\text{eq}}) + \rho C_p \text{VPD}}{\gamma \text{I}_{\text{eq}}} - 1 \right].
\]
where $R_{aero}$ is the aerodynamic resistance ($=1/ku$, in $s\ m^{-1}$); $k$ the von Karman’s constant ($=0.4$); $u_*$ the friction velocity (in $m\ s^{-1}$); $Z$ the reference height ($=2.2\ m$); $d$ the zero plane displacement ($=0.63\ \times\ \text{canopy height}$); $Z_0$ the roughness length ($=0.13\ \times\ \text{canopy height}$ (in m); Montieth, 1973); $E$ the latent heat flux (in $W\ m^{-2}$); $\rho$ the air density (in $kg\ m^{-3}$); $C_p$ the specific heat of air at constant pressure (in $J\ kg^{-1}\ C^{-1}$); and $VPD$ the vapor pressure deficit (in hPa).

3. Results

3.1. Microclimate and vegetation condition

The PPFD reached a maximum of 63.9 mol $m^{-2}$ per day in late June and then decreased gradually (Fig. 1a). The daily average air temperature and soil temperature ranged from 2 to 15.5 and 6.3 to 19.2 $^\circ C$, reaching the maximum of 15.5 and 19.2 $^\circ C$, respectively, in mid July (Fig. 1b and c). Rain fell periodically in both
2002
May Jun Jul Aug Sep Oct
DOY

Fig. 2. Daily moisture and canopy physiological conditions for days of the year 140–280 in 2001 (left-hand side) and 2002 (right-hand side) in Haibei, Qinghai, China. (a) Ratio between latent heat flux ($l_E$) and equilibrium evaporation ($l_{E\text{ eq}}$); and (b) Penman-Monteith canopy resistance ($R_{\text{canopy}}$). Gray zones correspond to soil water content shown in Fig. 10.

The closure of the surface energy budget was examined by performing a linear regression between the sum of eddy fluxes ($H + l_E$) and the available energy ($R_{\text{a}} - G$):

$$ (H + l_E) = 0.679 \times (R_{\text{a}} - G) - 0.174, \quad r^2 = 0.89 $$

(3)

where $H$ is the sensible heat flux and all the flux values were daily averages (in MJ m$^{-2}$).

On 20 August 2001, the measured LAI and aboveground biomass were 3.1 and 347 g d.w. m$^{-2}$, respectively, which were the annual maxima. In 2002, the LAI began to increase from late May and reached a maximum of 3.8 on 16 July (DOY197) and then decreased slowly (Fig. 3). The aboveground biomass reached a maximum of 283 g d.w. m$^{-2}$ on 30 July (DOY211), remained high during August, then decreased rapidly in September.

3.2. Diurnal courses of carbon dioxide exchange

During 9–15 August 2001 (DOY221-227), the weather was fairly clear. For several hours after midnight (Beijing Standard Time), $F_{\text{CO}_2}$ was positive (i.e. emission) and then became negative (i.e. uptake) at 08:00 (Fig. 4). The magnitude of $F_{\text{CO}_2}$ increased as PPFD increased, to a maximum uptake of $-12.4$ μmol m$^{-2}$ s$^{-1}$ at 13:00 h. Then, CO$_2$ uptake started to decrease until the grassland became a net source of CO$_2$ around 20:00h. Typically, $T_{\text{air}}$ reached a minimum of 2.2 °C at 07:00 h and a maximum of 19.9 °C at 17:00 h, indicating a large diurnal amplitude of air temperature variation. $T_{\text{soil}}$ reached a minimum of 12.4 °C at 08:00 h and a maximum of 20.6 °C at 18:00 h. The values of SWC were between 0.267 and 0.286 because of daytime evapotranspiration and small precipitation on the morning of 12 August. The diurnal pattern of VPD closely followed that of $T_{\text{air}}$ and reached a maximum of 15.4 hPa at 17:00 h. $R_{\text{canopy}}$ remained low at 100–140 s m$^{-1}$ in the forenoon, but started to increase in the
Fig. 3. Mean and standard deviation (N = 5) of the aboveground biomass and the leaf area index for the *K. humilis* alpine meadow in 2002.

afternoon and reached a maximum of 168 s m$^{-1}$ around 20:00 h.

We examined the seasonality of the diurnal course of \( F_{\text{CO}_2} \) (Fig. 5). The maximum CO$_2$ uptake of $-10.8 \mu$mol m$^{-2}$ s$^{-1}$ occurred around 13:00 h in August 2001. The maximum CO$_2$ release was 4.4 $\mu$mol m$^{-2}$ s$^{-1}$ at 01:00 h in August 2002. The amplitudes of \( F_{\text{CO}_2} \) were large in July and August when aboveground biomass peaked, but small in May–June and September. The time courses of \( F_{\text{CO}_2} \) in August and September of 2001 were similar to those in 2002, but their magnitudes were larger in 2001. It was noted that the light intensity was similar between August 2001 and August 2002.

To investigate the relationship between PPFD and \( F_{\text{CO}_2} \), we plotted the \( F_{\text{CO}_2} \) of Fig. 5 against PPFD (Fig. 6). The magnitude of \( F_{\text{CO}_2} \) increased as PPFD increased every month. At the same level of PPFD, particularly in July–August, the afternoon magnitudes of \( F_{\text{CO}_2} \) was smaller than those in the forenoon.

### 3.3. Seasonal changes in carbon dioxide exchange

The net ecosystem CO$_2$ production for daytime (NEP$_{\text{daytime}}$), nighttime (NEP$_{\text{nighthime}}$) and daily period (NEP$_{\text{total}}$) were computed by integrating values of CO$_2$ uptake (i.e., $-F_{\text{CO}_2}$) for the corresponding periods (Fig. 7). In 2002, NEP$_{\text{daytime}}$ started to increase from late May, and reached its maximum around 30 June (DOY181) with a local maximum of 5.4 g C m$^{-2}$ per day on 7 July. NEP$_{\text{daytime}}$ started to decrease from August until late September when the values dropped as low as those observed in May. NEP$_{\text{nighthime}}$ showed the opposite change, reaching a minimum of $-2.5$ g C m$^{-2}$ day on 19 August (DOY231). NEP$_{\text{total}}$ showed similar changes to those of NEP$_{\text{daytime}}$ and peaked at 3.9 g C m$^{-2}$ per day on 7 July (DOY188). NEP$_{\text{total}}$ rarely became negative, but reached a minimum of $-1.3$ g C m$^{-2}$ per day on 29 September (DOY272). The cumulative NEP$_{\text{total}}$ for the growing season from 23 May to 30 September 2002 was 153.1 g C m$^{-2}$. We performed a linear regression (Fig. 8) of NEP$_{\text{daytime}}$ (Fig. 7) on integrated PPFD. The slopes of the regression line showed similar seasonal patterns to those of LAI.

To investigate the relationship between ecosystem respiration and temperature, the nighttime averaged \( F_{\text{CO}_2} \) was plotted against the nighttime averaged soil temperature (Fig. 9). The nighttime \( F_{\text{CO}_2} \) increased exponentially with soil temperature. However, for a narrow range of soil temperatures in each month, such relationship did not hold and the nighttime \( F_{\text{CO}_2} \) scattered all over. Next, to assess the effect of SWC on ecosystem respiration, the nighttime \( F_{\text{CO}_2} \) was plotted against SWC for five periods of
Fig. 4. Diurnal courses of hourly mean CO₂ exchange flux and environmental conditions on 13 August 2001. (a) Net carbon exchange rate ($F_{\text{CO}_2}$, □); incident photosynthetic photon flux density (PPFD, ●). (b) Air temperature ($T_{\text{air}}$, ○); soil temperature ($T_{\text{soil}}$, □); soil water content (SWC, △). (c) VPD (-); Penman-Monteith canopy resistance ($R_{\text{canopy}}$, ■); latent heat flux (lE, ▲).

4. Discussion

4.1. Carbon dioxide uptake potential

4.1.1. Diurnal amplitude

We compared the maximum CO₂ uptake (~10.8 μmol m⁻² s⁻¹; Fig. 5) at our site with those at other sites located in similar latitudes (Table 1). Our maximum

low SWC after heavy rainfall (Fig. 10). The CO₂ efflux seemed to increase as SWC decreased, and a CO₂ release of at least 0.94 μmol m⁻² s⁻¹ was observed under saturated conditions. For the three periods in 2002 (i.e., 23 May–1 June, 11–19 August, 24–31 August in Fig. 10), the nighttime averaged soil temperature increased (by 2.8–4.2 °C) as SWC decreased.
CO₂ uptake was two-thirds of those in a C₃/C₄ prairie in Oklahoma (−15.5 μmol m⁻² s⁻¹; Sims and Bradford, 2001) and a subalpine conifer forest of fir and black spruce in Colorado (−15.5 μmol m⁻² s⁻¹; 3050 m; Monson et al., 2002). But it was 20–40% less than those in C₃ prairies in Kansas (−25 μmol m⁻² s⁻¹; Ham and Knapp, 1998), a tallgrass prairie in Oklahoma (−31.8 μmol m⁻² s⁻¹; Suyker and Verma, 2001), and a Japanese C₃/C₄ grassland (−56.7 μmol m⁻² s⁻¹; Li and Oikawa, 2001). Similarly, we compared the maximum rates of CO₂ release (4.4 μmol m⁻² s⁻¹; Fig. 5) at our site with other sites (Table 1). It was 20–50% less than those in a subalpine conifer forest in Colorado (8–9 μmol m⁻² s⁻¹; Monson et al., 2002), C₄ prairies in Kansas (10 μmol m⁻² s⁻¹; Ham and Knapp, 1998).
Fig. 7. Changes in daily net ecosystem production in 2001 (upper graph) and 2002 (lower graph). NEP<sub>daytime</sub>, NEP<sub>nighttime</sub>, and NEP<sub>total</sub> are the sums of the net CO<sub>2</sub> uptake flux (−<i>F</i><sub>CO2</sub>) for daytime, nighttime, and all day, respectively.

Fig. 8. Linear regression of daytime accumulated CO<sub>2</sub> uptake flux (NEP<sub>daytime</sub>) on incident photosynthetic photon flux density. The regression follows a linear relationship: NEP<sub>daytime</sub> = a × PPFD. Monthly values are presented as follows: month (symbol, r<sup>2</sup>)—August 2001 (●, 0.0793, 0.418**), September 2001 (△, 0.0556, 0.324**), May 2002 (●, 0.00671, 0.0653 n.s.), June 2002 (●, 0.0398, 0.170*), July 2002 (×, 0.0773, 0.0441 n.s.), August 2002 (●, 0.0641, 0.222*), September 2002 (△, 0.0322, 0.0344 n.s.). The linear relationships were significant at **P<0.01, *P<0.05 levels of correlation coefficients. n.s. shows nonsignificant linear relationships.
Fig. 9. Relationship between nighttime CO₂ flux density and nighttime soil temperature in 2001 and 2002. Data show nighttime mean values. The exponential relation is described by $F_{\text{CO}_2} = 0.399 \exp(0.132 \times T_{\text{soil}})$, $r^2 = 0.394$, $P < 0.0001$.

Thus, this Tibetan alpine meadow ecosystem has lower potential of CO₂ uptake and release than C₄ grasslands, but similar potential of net CO₂ uptake to alpine conifer forest because of its equivalent

Fig. 10. Relationship between nighttime CO₂ flux density and nighttime soil water content during a dry spell after heavy rain (1, 18–28 August 2001; 2, 23 May–1 June 2002; 3, 7–18 June 2002; 4, 11–19 August 2002; 5, 24–31 August 2002). Data show daily mean values. In 2002 during 23 May–1 June, 11–19 August, and 24–31 August (the 2nd, 4th, and 5th periods, respectively), daily mean nighttime soil temperature rose gradually (first day–last day, °C): 1, 11.7–11.5; 2, 7.5–11.7; 3, 13.5–15.7; 4, 12.9–15.7; 5, 12.6–15.8. The linear relation is described by $F_{\text{CO}_2} = -9.97 \times \text{SWC} + 6.77$, $r^2 = 0.545$, $P < 0.0001$. The $K$. humilis meadow had relatively high LAI (∼3.1) with low aboveground biomass (ca. 300 g m⁻²). With a similar LAI to our study site, the aboveground biomass in a grassland in Oklahoma was reported to be as high as 800 g m⁻². The lower ratio of aboveground biomass to LAI in the $K$. humilis meadow might have been
Table 1

<table>
<thead>
<tr>
<th>Site</th>
<th>Latitude, longitude</th>
<th>Altitude (m a.s.l)</th>
<th>Maximum LAI</th>
<th>Maximum aboveground biomass (g d.w. m$^{-2}$)</th>
<th>Maximum CO$_2$ uptake (µmol m$^{-2}$ s$^{-1}$)</th>
<th>Maximum CO$_2$ release (µmol m$^{-2}$ s$^{-1}$)</th>
<th>Maximum NEP total (g C m$^{-2}$ per day)</th>
</tr>
</thead>
<tbody>
<tr>
<td>C$_3$ alpine meadow$^a$</td>
<td>37°37’N, 101°18’E</td>
<td>3250</td>
<td>3.8</td>
<td>283</td>
<td>$-$10.8</td>
<td>4.4</td>
<td>3.9</td>
</tr>
<tr>
<td>C$_3$/C$_4$ prairie$^b$</td>
<td>36°36’N, 99°35’W</td>
<td>630</td>
<td>1.5</td>
<td>370</td>
<td>$-$15.5</td>
<td>$-$</td>
<td>$-$</td>
</tr>
<tr>
<td>Subalpine conifer forest$^c$</td>
<td>40°02’N, 105°32’W</td>
<td>3000</td>
<td>4.2</td>
<td>$-$</td>
<td>$-$15.5</td>
<td>8.0–9.0</td>
<td>1.0</td>
</tr>
<tr>
<td>C$_3$ prairie$^d$</td>
<td>39°12’N, 96°35’W</td>
<td>324</td>
<td>1.6</td>
<td>414</td>
<td>$-$25.0</td>
<td>10.0</td>
<td>4.9</td>
</tr>
<tr>
<td>C$_4$ prairie$^e$</td>
<td>39°03’N, 99°32’W</td>
<td>441</td>
<td>3.2</td>
<td>1100</td>
<td>$-$</td>
<td>$-$</td>
<td>6.3</td>
</tr>
<tr>
<td>Tallgrass prairie$^f$</td>
<td>36°56’N, 98°41’W</td>
<td>$-$</td>
<td>2.8</td>
<td>$-$</td>
<td>$-$31.8</td>
<td>11.4</td>
<td>8.4</td>
</tr>
<tr>
<td>C$_4$/C$_3$ grassland$^g$</td>
<td>36°06’N, 149°06’E</td>
<td>27</td>
<td>5.5</td>
<td>$-$</td>
<td>$-$56.7</td>
<td>21.6</td>
<td>$-$</td>
</tr>
</tbody>
</table>

$^a$ Qinghai–Tibetan Plateau, China (this study).

$^b$ Oklahoma, USA (Sims and Bradfield, 2001).

$^c$ Colorado, USA (Monson et al., 2002).

$^d$ Kansas, USA (Ham and Knapp, 1998).

$^e$ Kansas, USA (Kim et al., 1992).

$^f$ Oklahoma, USA (Suyker and Verma, 2001).

$^g$ Tsukuba, Japan (Li and Oikawa, 2001).
4.1.2. Seasonal maximum

Sequestered a large amount of CO2 (153 g C m\(^{-2}\)) uptake potential. Consequently, the alpine ecosystem in Kansas and Oklahoma, USA (4.9 – 8.4 g C m\(^{-2}\) per day) reached a fairly high value under clear weather conditions in August 2001. One of the reasons could be the low nighttime temperature (e.g. near freezing air temperature 2.2 °C) that limited CO2 emission. The effect of SWC on CO2 influx in the alpine ecosystem seems different from that reported for relatively dry grassland (Hunt et al., 2002), likely because of the different range of SWC encountered. However, the decrease of F\(_{CO2}\) with SWC in Fig. 10 could be an artifact of changing soil temperatures whose effect was not separated in this analysis, and therefore deserves further investigation.

5. Summary and conclusion

Considering the vast area of about 2.5 × 10^6 km\(^2\) of the Qinghai-Tibetan Plateau with the large variation in topography, it seems difficult to determine a ‘representative’ site for CO2 flux observation. Our study site is located at relatively low altitude within the vast plateau with averaged elevation of above 4000 m. However, the alpine Kobresia meadow ecosystem is one of the most widely distributed vegetation types on the Qinghai-Tibetan Plateau (Zhou, 2001). Such alpine meadows occur in areas with elevation ranging from 3200 to 5200 m with variable climatic conditions. If we are justified in assuming that carbon budgets depend on more on vegetation type than on any particular geological conditions, then the knowledge obtained from the current ecosystem should provide important insights into our understanding of the carbon dynamics of the grassland ecosystems on the Qinghai-Tibetan Plateau. It should be noted that various grassland ecosystems occupy more than 60% of the plateau (Wang et al., 2002).

We measured the CO2 flux by the eddy covariance method for two growing seasons in an alpine Kobresia meadow on the Qinghai-Tibetan Plateau. These measurements provided the first CO2 exchange data in an extreme environment from China for a blank
area of the global flux network. The evidence from the observations led to the conclusion that the alpine meadow was a CO$_2$ sink, at least during the two growing periods sampled. The current CO$_2$ sink strength seems small and is comparable with many other sub-alpine ecosystems reported so far. Moreover, temperature was the major environmental control on CO$_2$ exchange; evidently, low temperature limited ecosystem respiration. Soil water availability was high in the alpine meadow. The high soil water might reduce the ecosystem respiration. It seems to be difficult to separate the effects of the temperature and soil water content on the ecosystem carbon budget. However, soil water content varied sharply during periods as short as a week or so, while temperature varied less and more gently at similar temporal scales. We, therefore, tentatively conclude that high soil water content is correlated with low ecosystem respiration at nighttime for this alpine ecosystem.

The alpine meadow exhibited a relatively low ecosystem CO$_2$ uptake, considering the fairly high leaf area index of about 3. From this study, we further conclude that the alpine meadow has the potential to sequester carbon, but that potential appears to be small because of the limitation of low temperature. Further studies are needed to clarify the environmental controls on the carbon dynamics of alpine meadow ecosystems.

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