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Inter-annual carbon dioxide uptake of a wet sedge tundra ecosystem in the Arctic

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ABSTRACT

The CO_2 flux of a wet sedge tundra ecosystem in the Arctic, at Barrow, Alaska, has been measured by the eddy correlation method since spring 1999, and the CO_2 uptake by the vegetation during the spring and growing periods was examined between 1999 and 2000. CO_2 flux changed to a sink immediately after the spring thaw in 1999 and the photosynthetic activity was high in the first half of the growing period. At this time the air temperature was low and solar radiation was high. In the 2000 season, the temperature was approximately 5 °C lower during the snow-covered period, and increased up to 5 °C higher right after the spring thaw but the solar radiation decreased to two thirds of that in 1999. Thus, we found different CO_2 accumulation during the snowmelt and the following two weeks between both years. The difference in the climate at beginning shoulder period of the growing season resulted in the difference of CO_2 accumulation through the growing period. The maximum level of photosynthetic potential ($P_{\max}$) in late July was analyzed as being almost the same at 20 $gCO_2\ m^{-2}\ d^{-1}$ for both years. However, the weekly average peak CO_2 uptake was 16.4 and 11.9 $gCO_2\ m^{-2}\ d^{-1}$ in 1999 and 2000, respectively, with the lower number in 2000 caused by the low radiation with high air temperatures. The CO_2 accumulation during the spring and through the growing periods was a net sink of 593 $gCO_2\ m^{-2}$ in 1999 and a sink of 384 $gCO_2\ m^{-2}$ in 2000. High CO_2 accumulation in 1999 was caused by earlier development of the vegetation, and the lower CO_2 uptake in mid summer in 2000 was caused by unseasonable weather.

1. Introduction

Obvious warming that affects global climate has been observed in the Arctic (Martin et al., 1997; Rigor et al., 2000) and most of the computer simulations predict that the Arctic will be warmed by an increase of atmospheric CO_2 (e.g. Cubasch and Meehal, 2001). Tundra ecosystems contain about 14% of the total terrestrial carbon pool (Post et al., 1982; Gilmanov and Oechel, 1995) accumulated over the Holocene. There has been a significant shift in the Arctic CO_2 flux from a sink to a source with warming in the mid 1970s to mid 1990s (Coyne and Kelly, 1975; Oechel et al., 1993; Oechel and Vourlitis 1994; Oechel et al., 1995; 2000), suggesting that the pool of carbon in the Arctic is highly sensitive to climate change. Also, the high-latitude warming is expected to increase greenhouse-gas emission, which can accelerate global warming (Oechel and Vourlitis, 1997; Oechel et al., 2000). Thus, future changes in climate and ecosystem greenhouse fluxes are difficult to predict. Despite the obvious importance of Arctic ecosystems and the field research that has been done, we are still unable to state the current or future greenhouse gas budgets and the relationships between climate change and net ecosystem fluxes. Therefore, it is still important to understand the relationship between the climate and ecosystem's dynamics, and to determine the CO_2 budget in Arctic ecosystems under the current conditions.

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We have obtained CO_2 flux from the wet tundra ecosystems on the Northslope of Alaska during previous growing seasons (Yoshimoto et al., 1996; 1997; Harazono et al., 1998; 2000; Vourlitis et al., 2000) and found yearly differences in CO_2 exchange. In order to determine the annual carbon budget of tundra ecosystems, winter flux measurements are important (Oechel et al., 1997; Mast et al., 1998; Fahnestock et al., 1999), especially in the Arctic, where the winter/cold period is longer than eight months. However, there are few winter CO_2 exchange results in the Arctic because of difficulties in the measurements during the winter weather. We had been trying to measure continuous fluxes in the Arctic, but had failed for the past two seasons due to unexpected problems, such as cables being chewed by foxes, disconnection of cables due to low temperatures, sensor icing, a sampling tube blockage by icing, etc. However, we did obtain a fragmented CO_2 flux in the Arctic during the cold spring and after the 2000–2001 season.

This paper reports the annual CO_2 uptake measured at the ecosystem level for the coastal tundra ecosystem at Barrow, Alaska. The field measurements have been carried out since the snow-season in 1999 and are continuing to present. The goals of this study are (1) to examine the seasonal variation of CO_2 flux and its inter-annual difference, (2) to reveal the climatic effect on the CO_2 budget, and then (3) to quantify the CO_2 budget of Arctic coastal tundra ecosystems under current weather conditions.

2. Site and flux measurement methods

CO_2 flux, energy flux and meteorology have been measured since spring 1999 at the coastal wet sedge tundra at Barrow, Alaska. The site (71°19′12.5″N, 156°37′20.211″W, elevation 1 m) is located near the National Oceanic and Atmosphere Administration's (NOAA) Climate Monitoring and Diagnostics Laboratory (CMDL) located at the northern edge of a dried lagoon called the Central Marsh. The fetch is more than 0.5 km with the prevailing wind. Vegetation consists of wet sedges, grasses, moss and lichens, and the dominant species is *Arctophila fulva* (Walker and Acevedo, 1987; Oechel et al., 1995). The vegetation was completely flooded after snow melt. The growing season starts mid- to late June and senescence occurs at end of August to late September depending on the weather in a given year (Oechel et al., 1995; Oechel and Vourlitis, 1997). Prevailing winds at the site were predominantly easterly, and westerly some of the time. The hourly average wind speed at a height of 1.9 m was over 3 m s^{-1} during most of the growing season. Southerly winds average around 2 m s^{-1}, but are infrequent. A 100 m boardwalk from the nearest bank at the edge of the dried lagoon to the measurement point (equipment box) and an additional 30 m removable extension to the measurement tower were installed in April 1999. We used line power with a voltage stabilizer at the site; thus no advection of generator exhaust could affect the measurements.

Fluxes of heat, water vapor, momentum and CO_2 were measured at a height of 1.9 m above the ground or 1.6 m over snow cover using an eddy correlation system with a three-dimensional sonic anemometer (Kaijo, DA-600) and an open-path infrared gas analyzer (IRGA; Advanet, E009a). The path length of each sensor was 0.1 m and the sensor separation distance between the center of sonic anemometer and open-path IRGA was 0.16 m. Three components of wind speed, air temperature, CO_2 and water vapor concentration from the above sensors were sampled at 10 Hz over a 30 min interval and recorded on a magneto-optical (640 MB) disc by a digital recorder (Teac, DRM3). Each disk allowed for 57 d of continuous data collection.

Supporting micrometeorological data such as temperatures, humidity, radiation, wind speed, wind direction and soil heat flux were also measured at the observation tower (6.2 m tall). Two sets of ventilated platinum resistance temperature probes and moisture sensors (Vaisala, HMP45A) were set at heights of 1.2 and 3.2 m, and type-T thermocouples were used to measure the soil/water temperature at depths of 0.01, 0.05, 0.1, 0.2 and 0.4 m below the surface. A four-component (short- and long-wave radiation with both directions up and down) radiometer (Eko, MR-40) and Photosynthetic Active Radiation (PAR) sensors (Li-Cor, LI-190) were mounted on the tower at a height of 2.2 m for measurement of albedo and energy flux, and PAR, respectively. Additional light sensors (Kips & Zonen, PCM-3; Li-Cor, LI190) were set at the moss surface to measure the light condition within the vegetation or under the snow. The vegetation/snow surface temperature was measured from the measurement tower by an infrared thermometer (Minolta, model 505). Five sets of optical signal detection cup anemometers (Makino, AF-750) were set at heights of 0.5,1.2, 2.1, 3.2 and 4.6 m for wind profile measurements with a wind vane (Makino, VF16) set at 4.2 m. Precipitation was measured by a tipping-bucket rain gage (Texas Electronics, TE525MM). Soil heat flux was measured using six heat flow plates (Eko, MF-9) buried around the tower at a depth of 0.01 m beneath the moss.

Furthermore, the CO_2 concentration gradient between the two heights (1.2 and 3.2 m) was measured by a closed path IRGA (Li-Cor, LI6262) to examine the eddy CO_2 flux under unsuitable weather. The fluxes by the gradient measurements were determined by both an improved aerodynamic method (Harazono and Miyata, 1997; Miyata, 2001) and the standard aerodynamic method. The former was applied when the friction velocity determined by sonic anemometer was available, and the wind profile was applied for the latter.

Micrometeorological variables were sampled every 5 s through a multiplexer (Campbell, AM416), and the 15 min averages were recorded using data loggers (Campbell, CR23X and CR10X) and storage modules (Campbell, SM4M). After quality control of these obtained data, the 30 min averages were used for the analysis.

Calibration of the open-path IRGA was carried out in a laboratory using two levels of standard gas (around 350 and 480 ppm; the cylinder was replaced during the 2 yr). The sonic anemometer calibration was carried out at the field, in the intervals of once per month during the growing season. Temperature, radiation and humidity sensors were calibrated at the beginning of the season, and the instrumental correction factor was checked approximately every 3 months by inter-comparing sensor or comparisons with a calibrated sensor.

3. Data analysis

Fluxes of sensible and latent heat, CO_2, and momentum were calculated from 16 400 sample data for a 30 min interval according to normal covariance calculation procedures. In order to determine fluxes more accurately, the following corrections were applied. The humidity effect on the sonic thermometry (Kaimal and Gaynor, 1991) was applied for the evaluation of the sensible heat flux; effects of the path-length and sensor separation on the spectrum were taken into account in the high-frequency range of the fluxes (Moore, 1986), the air density effect caused by the counter-flow on latent and CO_2 fluxes (Webb et al., 1980; Leuning et al., 1982), and the effect of cross-sensitivity of water vapor and CO_2 on the CO_2 flux due to the optical filter characteristics of the E009a sensor (Leuning and Judd, 1996). Coordinate rotation (Tanner and Thurtell, 1969) was only applied when the right angles of the horizontal and vertical axes of the sonic anemometer were guaranteed by regular sensor maintenance.

Calculated fluxes obtained during rain, fog and low wind conditions were removed. Then data were checked and extreme fluctuations, i.e. high amplitude greater than three times the three-term average, were also removed. These removed data were caused by instrumental problems due to dew and snow on the open-path IRGA and sonic anemometer after rain, snow and foggy weather, and flow distortion by upwind equipment in a few cases. Further data gaps were caused by unexpected disconnection of a signal cable (under low temperature), power loss and damage to cables from animals.

Net ecosystem exchange, NEE, of a day is defined as the daily integration of measured CO_2 flux, fCO_2 and CO_2 storage term throughout the day as defined by eq. (1) (Fan et al., 1992; Wofsy et al., 1993; Falge et al., 2001). The storage term is the accumulated CO_2 within the air-mass between the measurement height, h and ground surface when the boundary layer was at a stable condition caused by poor turbulent mixing (Goulden et al., 1996). Here, ΔcCO_2 in the second term on the right-hand side of eq. (1) is the change rate of CO_2 concentration (mol m^{-3} h^{-1}) at a given height below the measurement height.

$$\mathrm{NEE} = \int_{\text{day night}} fCO_2 + \int_0^{24} \int_0^h \Delta cCO_2 \, dz \, dt. \tag{1}$$

The Barrow site was continuously windy; the hourly averaged wind speed was over 2 m s^{-1} even in the early morning. We could not detect obvious CO_2 concentration gradients and rapid concentration increase/decrease in the air layer below the flux measurement height. Furthermore, the flux measurement height was 1.9 m above the ground, which was lower than at other flux observation sites, because the sensor path length of the sonic anemometer and open path IRGA were as short as 0.1 m, allowing it to be close to the vegetation surface. Therefore, the CO_2 storage term at the site is negligible in calculations of NEE.

NEE is also related with ecosystem photosynthesis P, soil respiration R_{soil}, and aboveground respiration R_{plant} as defined by eq. (2) (e.g. Wofsy et al., 1993):

$$\mathrm{NEE} = R_{\text{soil}} + R_{\text{plant}} - P = R_{\text{eco}} - P. \tag{2}$$

In order to determine ecosystem respiration, $R_{(\text{eco})}$, from measured CO_2 flux, fCO_2, we divided $R_{(\text{eco})}$ into day and night as defined by eq. (3), which are shown with subscripts for day and night, respectively: It is difficult for our site to determine the ecosystem respiration at night according to previous wetlands studies (e.g. Whiting, 1994; Lafleur et al., 2001a) and a gap-filling strategy (Falge et al., 2001; Baldocchi et al., 2001), because Barrow is located at high latitude in the Arctic. There are few dark periods in the growing season. Therefore, the relationship between the measured CO_2 flux and the soil/water temperature was determined using data obtained during inactive periods of photosynthesis, such as mid-June, and at night after late-August (Mano et al., 2002). The relationship is applied to extrapolating daytime ecosystem respiration and that at night. Practical fitting to an exponential function is shown in the Results section.

$$\int_{\text{day}} f(\mathrm{CO_2})\,\mathrm{d}t = R_{\text{eco,day}} - P$$
$$\int_{\text{night}} f(\mathrm{CO_2})\,\mathrm{d}t = R_{\text{eco,night}}. \tag{3}$$

The flux data based on a 30 min average were quality controlled; as a result the data gaps were 39% and 52% during April and September in 1999 and 2000, respectively. Gap filling is recommended to examine the carbon budget from flux measurements (Falge et al., 2001). Missing or rejected data were filled using the above empirical relationship for R_{eco} and a simple model for CO_2 flux (Tundra Carbon Budget Model, TCBM). TCBM is a semi-empirical model classified as "Look-up tables" (Falge et al., 2001) and the "table" consists of several tables, each valid for a short period in order to fit the rapid changes during vegetation development. The parameters of basic processes in the model were fitted in each period by applying measured meteorological data, such as soil temperature, air temperature and PAR (Yoshimoto et al., 1997; Harazono et al., 1998).

The relationship between photosynthesis, P, and PAR is defined as eq. (4) (e.g. Wofsy et al., 1993; Lafleur, 1999): here, we replaced P by P_0 $f(x)$, a multiplication of potential photosynthesis, P_0 and a restraining function, $f(x)$, which reduces the photosynthesis by an unsuitable environmental factors, x, namely $0 < f(x) < 1$. The parameters α and $P_{\max}$ are the initial slope of the P_0–PAR relationship and hypothetical maximum of P_0, respectively. The model parameters were determined by approximately half-weekly periods in 2000 to reflect the rapid changes of the tundra vegetation (Mano et al., 2002).

$$P = f(x)P_0 = f(x)\frac{P_{\max}\,\alpha\mathrm{PAR}}{P_{\max} + \alpha\mathrm{PAR}}. \tag{4}$$

4. Results

4.1. Weather differences in two growing periods

Figure 1 shows the seasonal trends of 7-d-running mean of air temperature, T_a, solar radiation, R_s, and albedo, A_d, during two growing seasons in 1999 and 2000. The time series of albedo, A_d, showed clearly the first day of snow-free conditions (spring thaw) at the site (DOY 163 and 161 in 1999 and 2000, respectively). Continuous snow-cover in autumn in 2000 started 14 d earlier than that in 1999; thus the snow-free period, namely the vegetation growing period in 2000, was 12 d shorter. A_d was around 0.8 during the snow-covered period and was at a minimum of 0.08 after the spring thaw in mid-June. Then A_d increased gradually to about 0.18 with the growth of vegetation. The low A_d before the emergence of new leaves of the vegetation was thought to be caused by dark colored standing water at the site and the following increase was related to the development of vegetation. The maximum level of A_d within the growing period was higher in 1999 than in 2000.

During the snow-covered period until mid-June, R_s increased gradually with little difference between the two years. After spring thaw, there was an evident difference in R_s between the two years; it maintained higher levels during the following two weeks in 1999, while it decreased from a maximum level of 24 MJ m^{-2} d^{-1} to 9 MJ m^{-2} d^{-1} during the 3 wk following the spring thaw in 2000. After mid-July, R_s decreased gradually with the annual change of the incident angle of the sun. The daily amounts of R_s in the first half of August (DOY 210–225) were 20% lower in 2000 than in 1999.

Daily average air temperature, T_a during May (DOY 120–150) was lower in 2000, and increased to approximately 0 °C around the spring thaw (DOY 161–163) in both years. Then T_a increased rapidly up to 7 °C within a week in 2000 that was about 5 °C higher than in 1999 right after the spring thaw. The rate of air temperature rise before the spring thaw (DOY 135–160) was 0.27 °C d^{-1} in 1999 and 0.45 °C d^{-1} in 2000, even though there was little difference in solar radiation between both years. The wind speed during the same period was 4.18 m s^{-1} in 1999 and 3.60 m s^{-1} in 2000.

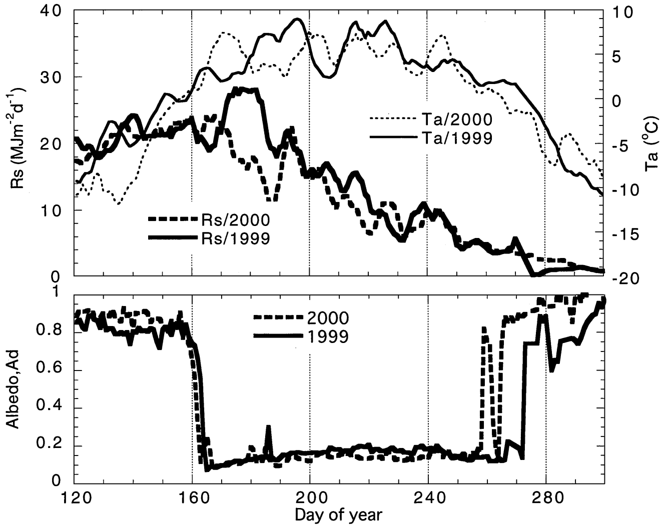

Fig. 1. Differences in the seasonal trends of air temperature, T_a solar radiation, R_s and albedo, A_d during the two growing seasons, 1999 and 2000. Measurements were continuous since April 1999, and T_a and R_s are shown as 7-d-running means of each daily average data set.

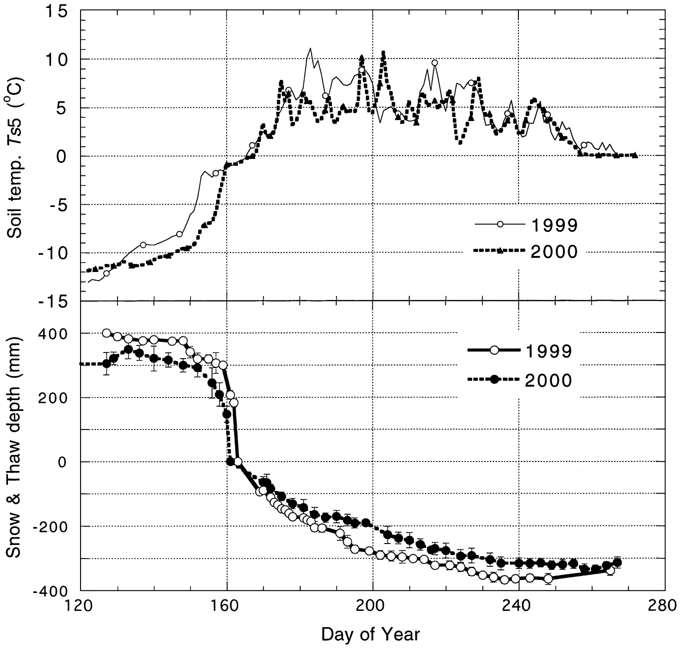

Fig. 2. Changes of seasonal trends of soil temperature at 0.05 m depth, T_{s5}, during the vegetation growing period in the both years (upper), and snow depth and thawed layer depth in 1999 and 2000 measurement season (lower).

Figure 2 shows soil temperature at 0.05 m, T_{s5}, snow depth and thawed layer depth in both years. Snow depth before spring thaw (DOY 120–160) was 0.05–0.1 m deeper in 1999, and soil temperature, T_{s5}, was 3–5 °C higher in 1999. After the snow-melt (DOY 160–180), thawing occurred more rapidly in 1999 in spite of little differences in T_{s5}. The maximum thaw depth was 0.37 m in 1999 and 0.34 m in 2000.

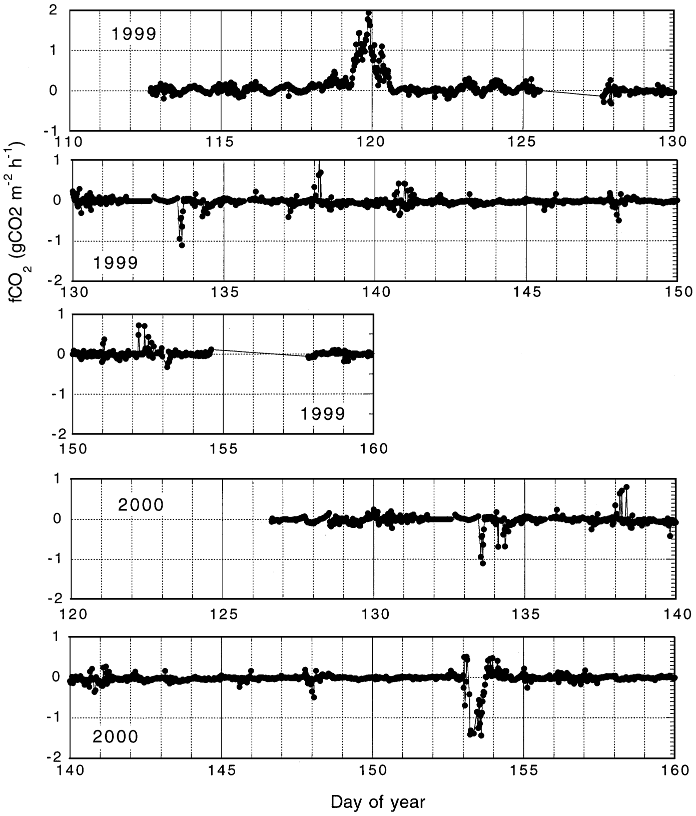

Fig. 3. Diurnal trends in ecosystem level CO_2 flux measured by eddy correlation technique at the Barrow site from snow-covered through snowmelt periods in 1999 and 2000. Each dot shows 30 min average. Negative value means downward CO_2 flux from the atmosphere to the ecosystem in relation to the CO_2 accumulation by photosynthesis and the positive means upward CO_2 flux.

Precipitation from June to September in 1999 and 2000 was 40.0 and 88.7 mm, respectively. The most rainfall (45.8 mm) in 2000 fell in July. The weather in 2000 was generally warmer in June and October, and cooler and rainier, with less solar radiation, in July than 1999.

4.2. *Daily variations of CO_2 flux*

Figure 3 shows the trend of CO_2 flux measured for the snow-covered period (until DOY 160 in each year). The positive values represent an upward (source) flux and the negative represents a downward flux (sink). Lack of data before DOY 126 in 2000 was caused by the disconnection of signal connector to the open path IRGA during a blizzard. Occasional large CO_2 fluxes were observed in this period, which were obvious upward (DOY 119–120, 1999) or downward (DOY 133, 1999 and DOY 153, 2000) and continued for a day or more. A large quantity of CO_2 efflux and downward fluxes occurred under high wind conditions. The efflux increased rapidly when the friction velocity, u_*, was over 0.25 m s^{-1}. The function was determined as eq. (5) (Harazono et al., 2000):

$$f\mathrm{CO}_2 = 0.089 + 46.9(u_* - 0.25)^2 \quad (R^2 = 0.978, n = 225). \tag{5}$$

The site was covered by approximately 0.38 and 0.33 m of snow during the episodes of large fluxes in each year. A snow saltation (saltation mass flux) occurred over 0.25 m s^{-1} of friction velocity (Pomeroy, 1989), which suggests the measured CO_2 flux related deeply with snow drift. Snow density measured at the site under cold period (below −15 °C) ranged

0.15–0.16 $g cm^{-3}$ in mid-May through May 26 and increased to 0.29 $g cm^{-3}$ on 4 June 2001. The densities during cold periods were lower than that of wind-packed snow density measured on Alaska arctic tundra (0.25–0.30 $g cm^{-3}$, Benson and Sturm, 1993). The low-density snow-pack layer at the site had the ability to contain more CO_2 and easily drifted. CO_2 concentration in the snow was higher than that in the air (e.g. Solomon and Cerling, 1987; Kim, 1998). CO_2-enriched air in the sparse snow-pack layer could be released into the air during blowing snow events when $u_* > 0.25$ m s^{-1}. Also, the CO_2 in the soil layer can be sucked into the air through plant structures by strong winds during snow drifting. The downward CO_2 flux during strong winds ($u_* > 0.25$ m s^{-1}) also increased quadratically with u_*, but we could not find the reason for the sink in this period. We will examine these occasional sinks in detail later, including additional measured data.

A small CO_2 sink was measured in the clear daytime (Fig. 3, DOY 113–124, 134–135, 143–144, in 1999 and DOY 129, 134–135, in 2000). CO_2 flux in 1999 showed a clear daily change of upward flux at night and downward flux in the daytime during DOY 113–118 and 122–124, while those in 2000 were unclear and the amplitude between daytime and at night was low. The downward CO_2 flux in midday in the tundra ecosystem was synchronised with daily variation of solar radiation when the air temperature was below –15 °C (Harazono et al., 2000). The effects of other meteorological parameters, such as wind speed and temperatures of air, soil and snow surface on the daily variation of sink CO_2 flux were examined, but no relationship was revealed for this period. A dynamic chamber measurement was applied to verify the sink activity during snow-covered period at the site in spring 2000, which also provided the sink CO_2 flux in midday (Harazono et al., 2000). However, the maximum amplitude of sink CO_2 flux by the chamber was one third smaller than that by eddy correlation measurement. The daytime sink flux during low temperatures over snow-covered tundra has to be examined in order to reveal the CO_2 budget during the winter–spring periods.

Trends of CO_2 flux during the growing period and after (starting at DOY 160) in 1999 and 2000 are shown in Figs. 4 and 5, respectively. CO_2 exchange during the spring thaw was different in both years. In 1999 there was almost no CO_2 flux on DOY 148–165 (Figs. 3 and 4); that changed to a significant downward CO_2 flux in the daytime and the amplitude of the CO_2 sink increased with each passing day after DOY 166. We could not find obvious CO_2 emissions before and after the spring thaw at the site in the 1999 season. In the 2000 season a weak source CO_2 flux was observed before and after the spring thaw (DOY 155–165), then the CO_2 efflux became evident in the daytime, and the daily cumulative CO_2 flux ranged to 2.8 gCO_2 m^{-2} d^{-1} emission during DOY 169–181. However, the efflux level during the snowmelt period was quite low at the flooded site compared to those reported for tundra vegetation in the Arctic (Vourlitis and Oechel, 1997; Vourlitis et al., 2000).

In mid-summer, daily variation of CO_2 flux showed a clear daytime sink and a low level of CO_2 emission at night. The peak level of CO_2 accumulation of over 2.5 gCO_2 m^{-2} h^{-1} appeared on DOY 201 in 1999 and approximately 2.0 gCO_2 m^{-2} h^{-1} on DOY 215 in 2000. Diurnal variation of CO_2 flux changed with varying solar radiation, temperature, humidity and wind speed. Scattered and fluctuating CO_2 fluxes were observed under weather conditions of rain, fog and high wind speeds over 10 m s^{-1}.

In the senescence period, the upward CO_2 flux at night became obvious, but the daytime sink CO_2 flux was observed through September. The sink strength during this period (DOY 235–244) was still high in 1999 and as a result, the daily CO_2 budget was negative until early September in 1999. After DOY 249 in 1999, we failed to obtain CO_2 flux data. In fall of 2000 there was unseasonable weather with frequent low clouds and thick fog, so most flux data were rejected. The daily variation of CO_2 flux in this period was not regular when compared to that in 1999, and the CO_2 accumulation during daytime was low. The change of daily cumulative CO_2 flux from sink to source occurred on DOY 247 in 2000, and the positive NEE increased to 2.9 gCO_2 m^{-2} d^{-1} on DOY 252. Because of the lack of data after DOY 249 in 1999 and data rejection in 2000, it was difficult to determine the exact end of the growing period, but CO_2 emission during the first half of September was higher in 2000 when compared to 1999.

4.3. Ecosystem respiration and NEE

The Arctic coastal wet sedge tundra ecosystem at Barrow showed different CO_2 accumulation and seasonally different characteristics under varying weather between 1999 and 2000. In order to reveal the vegetation response to the current weather, the ecosystem respiration and photosynthetic activity of the wet sedge

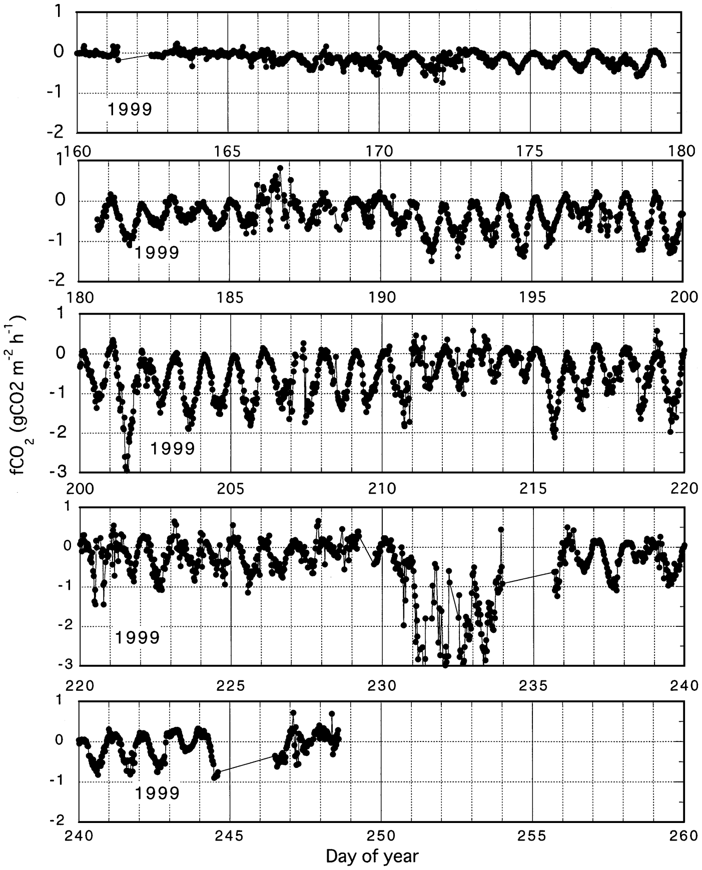

Fig. 4. Seasonal trends of diurnal changes of ecosystem level CO$_2$ flux at the site after snowmelt, growing and senescence period in 1999.

tundra was analyzed using simple parameterizations with empirical relationships.

Figure 6 shows the relationships between ecosystem level dark respiration, obtained from eddy correlation measurements during the dark period and that obtained when the photosynthesis was inactive, and soil temperature at 0.01 m, T_{s1}. Because of the long day-length at the high-latitude location, data available to examine the relationship were limited ($n = 70$ in 1999 and 56 in 2000), and the 30 min average fluxes were scattered. However, averages of 4–10 points of 30 min data within T_{s1} intervals provided clear relationships in 1999 and 2000, which were very close to each other. The exponential curve fit provided the following estimation of ecosystem respiration, $R_{eco,99}$ and $R_{eco,00}$ for each year:

$$R_{eco,99} = 0.0346 \exp(0.09755\ T_{s1}) \quad R^2 = 0.9543 \tag{6a}$$

$$R_{eco,00} = 0.0342 \exp(0.09439\ T_{s1}) \quad R^2 = 0.930 \tag{6b}$$

We assumed that the ecosystem respiration in daytime could be estimated by the same formula as was followed by other northern wetland studies (e.g., Fan et al., 1995; Lafleur, 1999; Griffis et al., 2000a; Lloyd, 2001; Lafleur et al., 2001a). The ecosystem respiration $R_{(eco)}$ was estimated at 30 min intervals, and the daily accumulation was calculated.

NEE is obtained from eq. (1) for the measured period, but the data gaps have to be filled to compare the inter-annual difference. Gap filling was carried out using eq. (2) with the empirical model of eqs. (4) and (6). In the model parameterization of TCBM, the

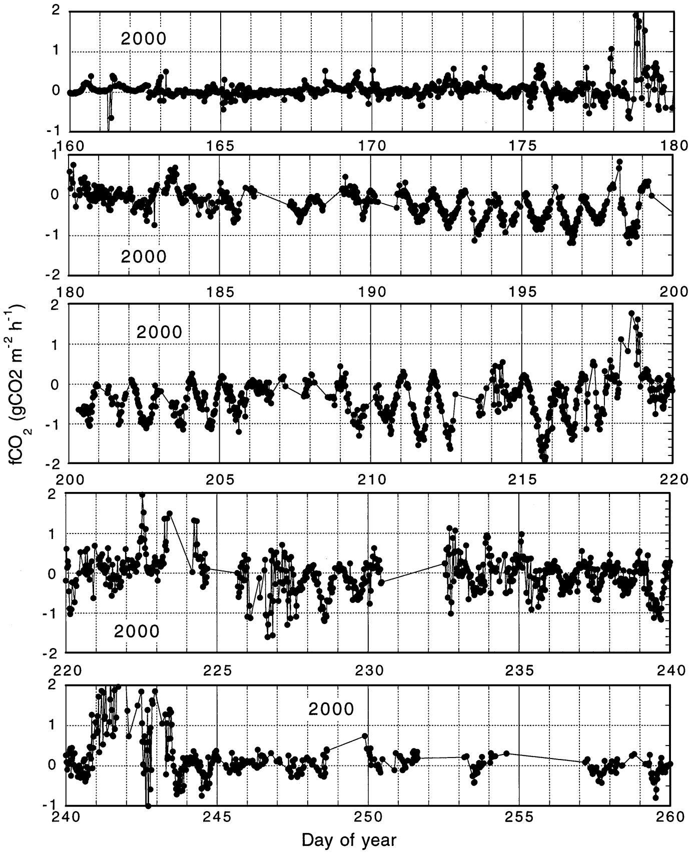

Fig. 5. Seasonal trends of diurnal changes of ecosystem level CO$_2$ flux at the site after snowmelt, growing and senescence period in 2000.

analysis term from 13 June to 21 September was divided into 24 periods, the lengths of which were different from the shortest of 3 d in the middle of June 2000 to the longest of 14 d in July–August 1999 (Mano et al., 2002). Here, P was determined for 30 min intervals by eq. (3), as the residual of measured $f(\mathrm{CO_2})$ and calculated R_{eco} by eq. (6), namely quasi-measured photosynthesis. Air temperature, T_{a}, was used as the variable of the restraining function, $f(T_{\mathrm{a}})$, in this paper. As a result, 24 sets of parameters were determined for 2000 and 10 sets for approximately 2 wk intervals in 1999. The small number of periods in 1999 was due to relatively stable weather and a small data gap. An example case of parameterization for the period between 25 and 28 July 2000 (DOY 207–210) is shown in Fig. 7.

The restraining function was fitted to a quadratic equation of air temperature, T_{a}, with peak temperature, T_{opt} as shown in Fig. 7. The presence of T_{opt} and midday suppression of P have been reported for vegetations in the Arctic (e.g. Tieszen, 1975; Oechel and Collins, 1976; Neumann et al., 1994; Whiting, 1995). At the Barrow site, parameterized T_{opt} varied from 1.07 °C in the thaw period to 13.9 °C in late July 2000 (Mano et al., 2002).

$P_{\max}$ showed seasonal variation ranging from 0.122 to 2.43 gCO$_2$ m^{-2} h^{-1} in 2000, which was higher in 1999. The biweekly averages of $P_{\max}$ and α during both growing periods in 1999 and 2000 are shown in Fig. 8. $P_{\max}$ at the beginning of the growing stage was higher in 1999 than in 2000, and that continued for the next 8 wk. α was almost constant until senescence for both years, and the difference was small. Initially, and following high levels of $P_{\max}$ in 1999, early sprout and development of the tundra vegetation was shown. The levels of $P_{\max}$ at the Barrow site are

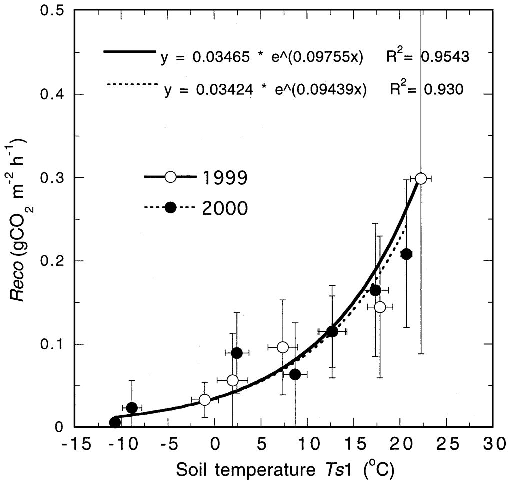

Fig. 6. Relationship of ecosystem respiration, $R_{(\mathrm{eco})}$, and soil temperature at 0.01 m depth T_{s1} in 1999 and 2000 growing periods. Plot data were obtained as 30 min average $\mathrm{CO_2}$ fluxes during dark periods and inactive periods of photosynthesis in middle of June and after late August.

higher than reported previously as 0.370–0.827 g $\mathrm{CO_2}$ $\mathrm{m^{-2}\ h^{-1}}$ by Whiting et al. (1992), 0.0367–2.23 $\mathrm{gCO_2}$ $\mathrm{m^{-2}\ h^{-1}}$ by Whiting (1994) and 0.911 $\mathrm{gCO_2\ m^{-2}\ h^{-1}}$ by Lafluer (1999). Since our fit to the photosynthesis–PAR relationship was for the potential photosynthesis, P_0 instead of P, the higher values in this study are reasonable.

Diurnal variations of calculated $\mathrm{CO_2}$ flux [$R_{\mathrm{eco}} + P_0 f(T_{\mathrm{a}}) = \mathrm{NEE}$] and measured $\mathrm{CO_2}$ flux are shown in Fig. 9 with PAR (dotted line) for the period between DOY 200 and 213, 2000. The relationship of $\mathrm{CO_2}$ flux between the calculated [$R_{\mathrm{eco}} - P_0 f(T_{\mathrm{a}})$] and the measured $f\mathrm{CO_2}$ for 2000 is shown in Fig. 10, which also shows the comparison between the TCBM calculation of photosynthesis, $-P_0 f(T_{\mathrm{a}})$ and the quasi-measured photosynthesis, $-P[=R_{\mathrm{eco}} - f\,\mathrm{CO_2}$ in daytime], which is the residual of measured $f\,\mathrm{CO_2}$ and calculated R_{eco}. The regression line between $-P_0 f(T_{\mathrm{a}})$ and quasi-measured $-P$ showed an underestimation of 20% for higher PAR ranges and an offset of -0.09 $\mathrm{gCO_2\ m^{-2}\ h^{-1}}$. Most data dropped less than 0.25 $\mathrm{gCO_2\ m^{-2}\ h^{-1}}$ and the standard deviation of the photosynthesis difference, $P_0 f(T_{\mathrm{a}})$ $-P$, was 0.170 $\mathrm{gCO_2\ m^{-2}\ h^{-1}}$ for all data ($n = 1735$) and 0.157 $\mathrm{gCO_2\ m^{-2}\ h^{-1}}$ for the range $-0.25 < P < 0$ ($n = 982$). The estimated values distributed uniformly around the regression line, so that the model estimation seems to reflect the photosynthesis–micrometeorology relationships. The practical error was smaller under the low PAR conditions, which was a major case for data rejection. Standard deviation for NEE [$R_{\mathrm{eco}} + P_0 f(T_{\mathrm{a}}) + f\,\mathrm{CO_2}$] was 0.156 ($n = 2243$). Data gaps of $f\,\mathrm{CO_2}$ and quasi-measured photosynthesis $-P$ were replaced by $R_{\mathrm{eco}} + P_0 f(T_{\mathrm{a}})$ and $P_0 f(T_{\mathrm{a}})$, respectively.

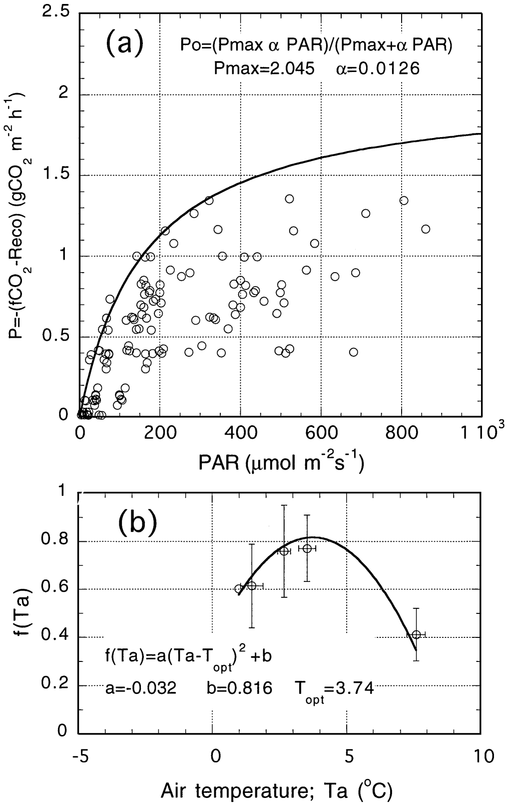

Fig. 7. An example of parameter $P_{\max}$ and α of the P_0–PAR relationship, and the restraining function $f(T_{\mathrm{a}})$ of the TCBM. α is the initial slope of the P_0–PAR relationship and the $P_{\max}$ is the hypothetical maximum of P_0. Actual photosynthesis P was defined as the multiplication of the potential photosynthesis P_0 and the restraining function $f(T_{\mathrm{a}})$, which is a function of air temperature, T_{a}, $0 < f(T_{\mathrm{a}}) < 1$.

Figure 11 shows weekly averages of NEE, R_{eco}, and $-P$ during the spring and growing period, in 1999 and 2000. The date on the horizontal axis in Fig. 11 shows the end day of the average period. Data gaps were filled for 30 min basis by the TCBM and eq. (6) shown above. Thus each daily cumulative was determined from the complemented data set; i.e. the NEE data consist of measured $f\,\mathrm{CO_2}$ and estimated $R_{\mathrm{eco}} + P_0 f(T_{\mathrm{a}})$, and P

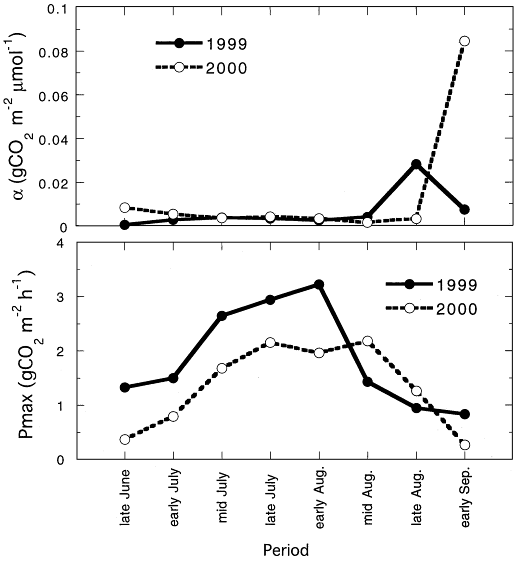

Fig. 8. Yearly difference of seasonal trends of $P_{\max}$ and α in 1999 and 2000, which were determined as parameters of potential photosynthesis of TCBM. Dots are biweekly averages determined during varied periods in each year.

consists of the quasi-measured photosynthesis and estimated $P_0 f(T_a)$, if there were gaps.

The seasonal trend of NEE is shown in Fig. 11. Weekly NEE indicates a small source before the snow-free period, but a negative NEE was observed during middle of May in both years (weeks through DOY 140 and 147 in Fig. 11). During the snow-free period, NEE showed different trends between 1999 and 2000. In 1999, NEE shifted from a source to a sink right after the spring thaw (week through DOY168, late June) and the sink strength increased to a weekly maximum level of -15.6 gCO_2 m^{-2} d^{-1} in late July (week through DOY 210). After the maximum activity period, it decreased with decreasing temperature and solar radiation (Fig. 1), and then NEE returned to positive on the week through DOY 259. In 2000, negative NEE occurred a week later than 1999. The smaller weekly maximum sink of -11.7 gCO_2 m^{-2} d^{-1} was observed in the same period (week through DOY 210). The weekly maxima of photosynthesis were 16.4 and 13.6 gCO_2 m^{-2} d^{-1} in 1999 and 2000 at the same period (week through DOY 210). After the peak season, photosynthesis decreased suddenly at the week through DOY 224 in 2000, when $P_{\max}$ was still high (mid-August in Fig. 8). The low levels of photosynthesis after the peak period in 2000 were attributed to low PAR and rainy weather.

During the end of the growing period, the weekly NEE was a source of 1.2 gCO_2 m^{-2} d^{-1} in week to DOY 252, 2000 (Fig. 11) in respect to higher air temperature (Fig. 1). In 1999 the NEE, partly including estimated photosynthesis, showed a sink during the same period and changed to a low source of 0.6 gCO_2 m^{-2} d^{-1} in the next week to DOY 259. Depending on the weather differences in each year, NEE during the senescence period showed different patterns. NEE changed to negative again in the week of DOY

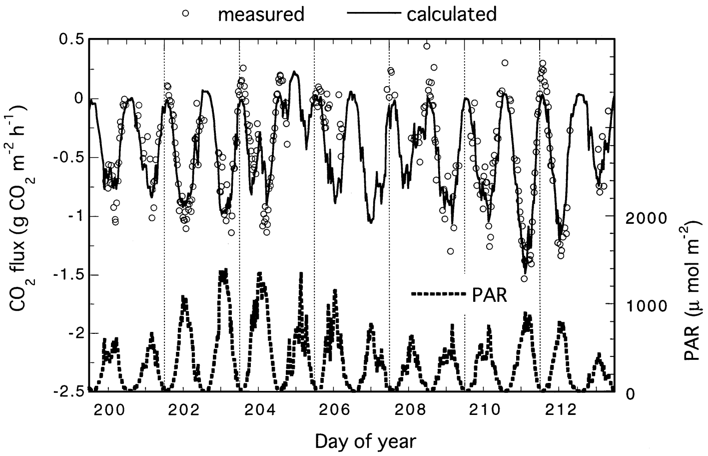

Fig. 9. Comparative diurnal variations of the estimated CO_2 flux by TCBM, $[-P_0 f(T_a) + R_{eco}]$, and measured CO_2 flux, f CO_2 (circles), under variable PAR (bottom of the figure) conditions during the middle to end July 2000.

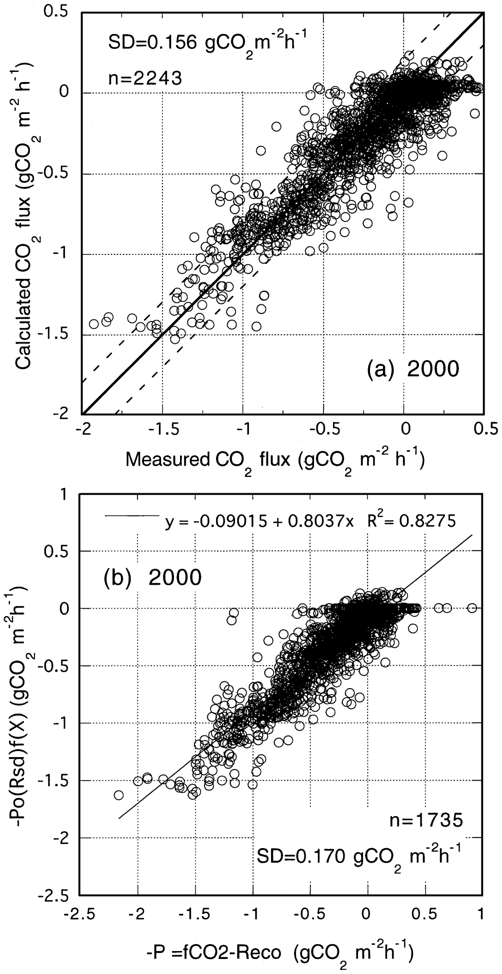

Fig. 10. Comparison of 30 min average CO_2 flux between estimated $[-P_0f(T_a) + R_{eco}]$ and measured (a), and that of photosynthesis between estimated by TCBM and quasi-measured, which was the residual of measured $f\,CO_2$ and estimated R_{eco} (b), for the growing period in 2000. SD denotes standard deviation.

259, 2000. In early October, the source NEE at the wet sedge tundra had almost the same CO_2 efflux of 0.7 $gCO_2\ m^{-2}\ d^{-1}$ in both years. The air temperature during this period was below −5 °C.

4.4. CO_2 uptake

Integration of NEE, R_{eco} and P were carried out from 30 min average data in which quality controlled data of fCO_2 were used first and the data gaps were filled from eqs. (4) and (6). Seasonal integration of NEE and R_{eco} during spring and growing periods show an accumulation of 593 $gCO_2\ m^{-2}$ of atmospheric CO_2 during the 21 wk (147 d) from May to September in 1999, and 384 $gCO_2\ m^{-2}$ during the same period in 2000. Estimated ecosystem respiration during the same period was 170 $gCO_2\ m^{-2}$ in 1999 and 157 $gCO_2\ m^{-2}$ in 2000. Seasonal accumulations of the quasi-measured photosynthesis [$f\,CO_2 - R_{eco}$ and $P_0f(T_a)$ filled] were 788 and 539 $gCO_2\ m^{-2}$ in 1999 and 2000, and seasonal accumulations of model estimation, $P_0f(T_a)$, were 762 and 579 $gCO_2\ m^{-2}$ in 1999 and 2000, respectively. The difference between the quasi-measured and estimated photosynthesis in each year was larger in 2000. The model estimation was lower than measured in 1999 but was larger in 2000. The difference was caused by the application of gap filling. $P_0f(T_a)$ had an offset providing negative values when the quasi-measured values were close to zero (Fig. 9), and most gap filling was applied for low PAR, especially in 2000. Also, the number of gap fillings shared a large portion in 2000. On the other hand, the calculated P values as residuals of eq. (2) for seasonal integration of NEE and R_{eco} were 763 and 541 $gCO_2\ m^{-2}$ in 1999 and 2000, respectively. Accumulations of the quasi-measured photosynthesis were close to those determined from eq. (2). If we use the quasi-measured photosynthesis to integrate the CO_2 budget, eq. (2) does not close in each year.

We tried to evaluate photosynthesis of a tundra ecosystem, which showed 3–7% differences in accumulated photosynthesis among gap-filling conditions. There were large differences in the cumulative carbon uptake by the wet sedge tundra vegetation between the two growing periods. This difference was due to the difference in CO_2 uptake by the tundra vegetation (early development, low accumulation in the latter half of the growing period) caused by different weather conditions.

5. Discussion

The CO_2 storage term evaluated previously for a bog peatland ecosystem showed considerable storage only during the hours around sunrise and sunset, but was negligible during daytime and even though the night (Lafler et al., 2001a). They estimated the storage term using the concentration difference measurements at two heights, and showed that the largest contribution of the storage term was 15% of eddy flux, but for only

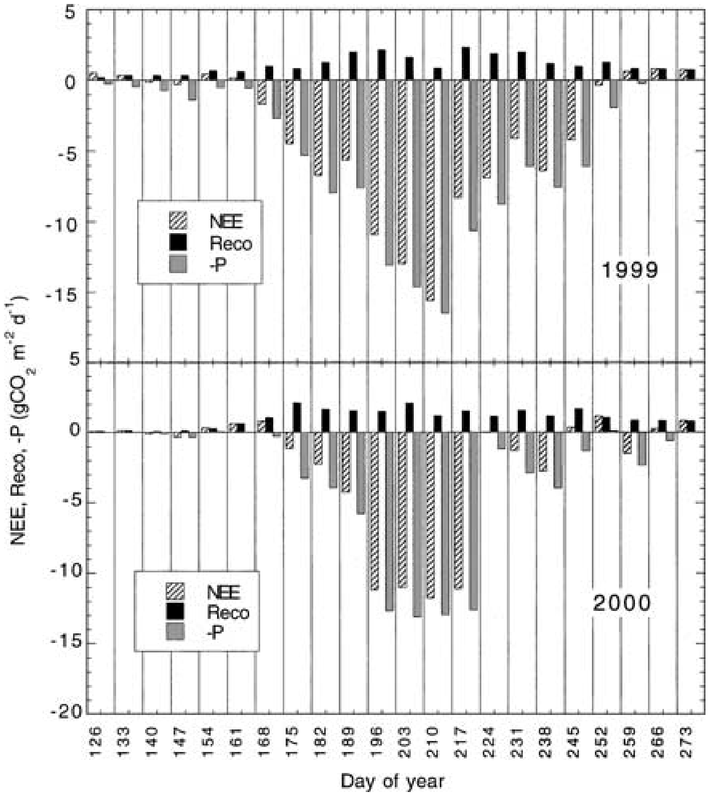

Fig. 11. Differences in seasonal trends of weekly averaged daily cumulatives of NEE, R_{eco}, and P in the 1999 and 2000 growing period. The day of year shows the end day of the average period.

a few hours a day. However, we measured few inversion conditions at the Barrow site. Furthermore, the air layer to be considered for the CO$_2$ storage was small because the eddy flux observation height was close to the surface. Thus, the contribution of the CO$_2$ storage term on NEE is thought to be negligible at the Barrow site, and the NEE determined from quality controlled CO$_2$ flux data is accepted as actual CO$_2$ exchange at the site, if there were few data gaps.

However, almost half of the flux data were either missing or rejected, thus the "Look-up tables" gap-filling technique (but for short terms) was applied in this study. The data available to analyze the relationship between ecosystem respiration, R_{eco}, and temperature were limited, and the relations with different soil depth and air temperature were examined. The scattered relationship of R_{eco} to temperature (Fig. 6) was caused by windy and poor vegetation coverage compared to other ecosystem types. Night-time CO$_2$ flux was affected by wind, and we found that the wind speed was related to the flux at the site (Mano et al., 2002) so that the CO$_2$ flux determined under $0.05 < u_* < 0.25$ m s^{-1} was used for fitting. Soil temperatures at 0.10–0.20 m have been widely used for wetland and forest ecosystems (e.g. Falge et al., 2001), but in Arctic tundra ecosystems, soil temperatures at deep levels changed too little to reveal the relationship. As a result, the soil temperature at 0.01 m depth, T_{s1} was used as the variable of the exponential function fit. The R^2 of the fit was relatively high, and the functions were not only close to each other in the two years but also close to the results for other northern wetlands (Fan et al., 1995; Lafleur, 1999; Griffis et al., 2000a; Lloyd, 2001; Lafleur et al., 2001). Thus, we state that the exponential function of T_{s1} is appropriate to estimate the ecosystem respiration.

CO$_2$ efflux before and right after the snow-free period have been widely observed at northern ecosystems (e.g. Lewis and Callaghan, 1976; Oechel et al., 1995; Vourlitis and Oechel, 1997; Vourlitis et al., 2000).

However, we could not observe evident CO_2 efflux during that period in 1999 (Ota et al., 2000). In 2000, as the level of CO_2 emission changed due to weather differences each day, small quantities of CO_2 were emitted during the same period (Fig. 11), but were low compared to previous results. We have another measurement site at moist tundra in Barrow approximately 1 km east from the wet sedge tundra site. We found clear differences in CO_2 efflux during the thawing period between both sites, and the seasonal accumulation of NEE was much higher at wet sedge tundra (personal communication from R. Zulueta and H. Kwon, 2002, in annual report to NSF). The difference could be explained by the presence of standing water at the wet sedge tundra site. Greenhouse gas flux studies on rice paddies have shown that standing water prevents gas exchange between below-ground and the air (Miyata et al., 2000; Leuning et al., 2000). Also, manipulation experiments that changed water tables of wet sedge tundra on the coastal plain near Prudhoe Bay showed an increase of ecosystem respiration by drained conditions (Oechel et al., 1998) and the other experiment on a moist tundra ecosystem at Barrow showed a clear decrease of ecosystem respiration by the high water table treatments (personal communication from G. Kinoshita and K.F. Huemmrich, 2002, in annual report to NSF). Thus, standing water at the site prevents the CO_2 emission from below ground to the atmosphere.

The inter-annual difference before the growing period in this study could be explained by different micrometeorology at the snow-covered vegetation. During the snow covered period, solar radiation and average wind speed showed little difference between 1999 and 2000, but the first day when the CO_2 flux changed to a sink and the following CO_2 exchange were different between both years (Figs. 8 and 11). Air and soil temperatures at 0.05 m depth were approximately 5 °C higher before the spring thaw in 1999. Higher soil temperature could bring about an earlier start of photosynthetic activity of moss, lichen and sedges under the snow-pack. The higher CO_2 sink during the snow-covered period in 1999 was also attributed to early vegetation growth, despite deeper snow depth in 1999.

Snow density measured at the site in spring 2001 was almost constant at 0.16 g cm^{-3} when the air temperature was below -15 °C; it became greater, at 0.29 g cm^{-3}, with increased air temperature. Snow density at Barrow in the cold season was quite low compared to other wind-packed Alaska Arctic tundra snow (Benson and Sturm, 1993), at densities of 0.25–0.30 g cm^{-3}, providing sparse snow structure. The sparse snow-pack layer in 1999 could enhance CO_2 diffusion between the vegetation surface and the atmosphere, while the relatively dense snow impedes CO_2 exchange. We did not measure the snow densities in spring 1999 and 2000, but we can estimate a sparse density in the first half of May 1999 and a higher density in May 2000 from temperature records. Thus, we have detected the smaller amplitudes of daily variation of sink–source pattern in 2000 compared to 1999, as shown in Fig. 3.

We found large differences in the seasonal CO_2 uptake (Fig. 11). Similar significant inter-annual variations of CO_2 exchange depending on weather conditions have been reported for other wetland and northern tundra ecosystems (e.g., Shurpali et al., 1995; Joiner et al., 1999; Fahnestock et al., 1999; Griffis et al., 2000a; Lafler et al., 2001a; 2001b). These studies concluded that the interannual differences were caused by differences in soil moisture and precipitation. Some of them also concluded that the climate around the shoulder periods of the growing season had an important role on the sink–source strength, but the relationship with micrometeorology was not made clear.

The present study was conducted on the flooded tundra ecosystem, so there was little seasonal change in soil moisture from the beginning of the growing season through to the end of the season; also we saw little difference in water tables between both years. However, we had quite a lower CO_2 uptake in 2000 than 1999; nevertheless we had greater precipitation in July 2000. Therefore, we state that the below-ground water content was not a major factor on CO_2 exchange in the wet sedge tundra in the Arctic.

Examination of micrometeorology and photosynthesis in this study revealed that higher soil temperature in the snow-covered period and weather conditions immediately after the spring thaw had important roles on the tundra vegetation development. PAR–photosynthesis relationships showed a clear difference between 1999 and 2000 (Fig. 8); the wet sedge tundra had clear photosynthetic activity immediately after snowmelt in 1999, and the continuously higher $P_{\max}$ in 1999 suggested that assimilated carbohydrate in the plants at the beginning allowed for faster development of the vegetation. Thus, the low air temperature and high solar radiation during the few weeks after the spring thaw in 1999 probably contributed to the faster development of the vegetation and higher CO_2 accumulation during the period. Weather conditions in the shoulder periods were important at northern wetlands

(Griffis et al., 2000a), but the initial shoulder is important in high arctic ecosystems.

Regarding the potential CO_2 uptake at the tundra ecosystem at Barrow, the maximum uptake of daily-integrated CO_2 flux was estimated by an empirical model as -20 gCO$_2$ m^{-2} d^{-1} in 1999, with -18 gCO$_2$ m^{-2} d^{-1} in 2000 (Mano et al., 2002). The potential photosynthetic activity of the tundra vegetation in mid summer was almost the same in both years. Thus, we conclude that high air temperature and low solar radiation after the snowmelt in 2000 resulted in a delay of the sprout of the vegetation, and the unsuitable weather in July and August 2000 decreased the CO_2 accumulation.

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