CO2 Exchanges between the Atmosphere and an Alpine Meadow Ecosystem on the Qinghai-Tibetan Plateau

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## Contents

<table>
<thead>
<tr>
<th>Section</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>Contents</td>
<td>i</td>
</tr>
<tr>
<td>Abstract</td>
<td>v</td>
</tr>
<tr>
<td>List of Tables</td>
<td>viii</td>
</tr>
<tr>
<td>List of Figures</td>
<td>ix</td>
</tr>
<tr>
<td>List of Symbols</td>
<td>xiv</td>
</tr>
<tr>
<td>Chapter 1 General Introduction</td>
<td>1</td>
</tr>
<tr>
<td>1.1. Background</td>
<td>1</td>
</tr>
<tr>
<td>1.1.1. Global climate change and terrestrial ecosystems</td>
<td>1</td>
</tr>
<tr>
<td>1.1.2. Measurement and modeling of carbon dioxide between the atmosphere and terrestrial ecosystems</td>
<td>2</td>
</tr>
<tr>
<td>1.1.3. An alpine meadow ecosystem on the Qinghai-Tibetan Plateau, China</td>
<td>4</td>
</tr>
<tr>
<td>1.2. Purpose of this study</td>
<td>5</td>
</tr>
<tr>
<td>1.3. Composition of the thesis</td>
<td>6</td>
</tr>
<tr>
<td>Chapter 2 Site description</td>
<td>8</td>
</tr>
<tr>
<td>Chapter 3 Diurnal changes of the CO2 exchanges</td>
<td>12</td>
</tr>
<tr>
<td>3.1. Introduction</td>
<td>12</td>
</tr>
<tr>
<td>3.2. Materials and Methods</td>
<td>12</td>
</tr>
<tr>
<td>3.2.1. Micrometeorology</td>
<td>13</td>
</tr>
<tr>
<td>3.2.2. Eddy covariance method</td>
<td>13</td>
</tr>
<tr>
<td>3.2.2.1. Fundaments of the eddy diffusion flux calculations</td>
<td>13</td>
</tr>
<tr>
<td>3.2.2.2. Measurement of eddy diffusion fluxes</td>
<td>19</td>
</tr>
<tr>
<td>3.2.2.3. Correction of measured turbulent flux data</td>
<td>19</td>
</tr>
<tr>
<td>3.2.2.4. Validity of measured flux data</td>
<td>21</td>
</tr>
</tbody>
</table>
3.2.3. Omega factor calculation .......................................................... 25
3.2.4. Bowen ratio calculation ............................................................ 27
3.2.5. Canopy resistance calculation ..................................................... 27
3.2.6. GPP and \( R_e \) calculations ....................................................... 27
3.2.7. LAI and biomass sampling ......................................................... 28
3.3. Results ....................................................................................... 28
  3.3.1. Micrometeorology and vegetation growth ............................ 28
  3.3.2. Sensible and latent heat fluxes .............................................. 29
  3.3.3. Carbon dioxide flux ............................................................... 30
  3.3.4. GPP and \( R_e \) ................................................................. 31
3.4. Discussion .................................................................................. 31
  3.4.1. Measurement accuracy of eddy covariance method ............ 31
  3.4.2. Environmental controls on the diurnal change of carbon dioxide dynamics ... 33
  3.4.3. Ecosystem carbon assimilation ability .................................... 34

Chapter 4 Seasonal changes of the CO\(_2\) exchanges ...................................................... 74
4.1. Introduction ............................................................................. 74
4.2. Materials and Methods ............................................................. 75
  4.2.1. NEP, GPP and \( R_e \) gap-filling methods ............................. 76
4.3. Results ..................................................................................... 77
  4.3.1. Micrometeorology ................................................................. 77
  4.3.2. Carbon dioxide flux ............................................................ 78
  4.3.3. GPP and \( R_e \) ................................................................. 79
  4.3.4. Annual sums of net ecosystem carbon dynamics ............... 81
4.4. Discussion ............................................................................... 81
  4.4.1. Environmental controls on the seasonal changes of net carbon dioxide exchanges ... 81
  4.4.2. Environmental controls on the seasonal changes of GPP and \( R_e \) ................. 82
  4.4.3. Ecosystem carbon assimilation ability .................................. 83

5.1. Introduction ...................................................................................................................... 102
5.2. Model description ........................................................................................................... 105
  5.2.1. Sim-CYCLE - Basic model- ......................................................................................... 105
    5.2.1.1. Overview of Sim-CYCLE ..................................................................................... 105
    5.2.1.2. Single-leaf processes ......................................................................................... 107
    5.2.1.2. Ecosystem-scale processes .................................................................................. 110
  5.2.2. Soil thermal profile ...................................................................................................... 117
    5.2.2.1. Thermal parameter .............................................................................................. 119
  5.2.3. Water budget .............................................................................................................. 120
    5.2.3.1. Rainfall, snow and thawing ................................................................................. 121
    5.2.3.2. Evaporation and transpiration ........................................................................... 121
    5.2.3.3. Runoff and Infiltration ....................................................................................... 126
5.3. Model experiment designs .............................................................................................. 127
5.4. Results ............................................................................................................................. 128
  5.4.1. Steady state experiment ............................................................................................. 128
  5.4.2. Transient experiment .................................................................................................. 129
  5.4.3. Model sensitivity analysis .......................................................................................... 130
5.5. Discussion ........................................................................................................................ 131
  5.5.1. Model accuracy .......................................................................................................... 131
  5.5.2. The Relationship between climate perturbations and ecosystem CO₂ exchanges ... 132
  5.5.3. The potential response of the ecosystem CO₂ exchanges against global warming.. 133

Chapter 6 General Discussion .............................................................................................. 165
6.1. Carbon dioxide assimilation capacity of alpine meadow ecosystem on the Qinghai-Tibetan Plateau .............................................................. 165

6.2. Environmental control on the CO₂ exchange on the different temporal scales ........ 166

6.3. The global warming effects on the ecosystem carbon sequestration ..................... 170

6.4. Recommendations for future research .................................................................. 171

Chapter 7 Conclusions ............................................................................................. 173

Acknowledgement .................................................................................................... 175

References ............................................................................................................... 177
Abstract

The Qinghai-Tibetan Plateau plays a critical role in regional and global climate change. More than 60% of the surface area of the plateau is covered by grasslands. To reveal the potential contribution of grassland ecosystems to climate change, I examined the CO$_2$ exchange between the atmosphere and an alpine *Kobresia* meadow (lat 37°29'-45°N, long 101°12'-23°E, 3250 m a.s.l.) on the northeastern Qinghai-Tibetan Plateau. The ecosystem CO$_2$ flux was measured continuously using the eddy-covariance method from August 2001 to December 2002. I examined the diurnal and seasonal variations of carbon fluxes and assessed their environmental and biological controls. I further explored the long-term pattern of carbon dynamics in the alpine ecosystem using meteorological and biomass data available for the period from 1981 to 2000 with a simulation model of the carbon cycle in terrestrial ecosystems (Sim-CYCLE).

The major findings from the CO$_2$ flux measurements were as follows: (1) The daily changes in CO$_2$ exchange showed CO$_2$ uptake in the daytime and CO$_2$ release in the nighttime in summer and small CO$_2$ release during all the day in winter. In 2002, net ecosystem productivity (*NEP*) during the five months of summer amounted to 138.4 g C m$^{-2}$. *NEP* during the rest of the year reached -59.9 g C m$^{-2}$ and the winter flux measurements provided very important information to estimate the annual carbon dynamics. The maximum daily CO$_2$ uptake (3.9 g C m$^{-2}$ day$^{-1}$) in this site was smaller than that in other grassland ecosystems at almost the same latitude, although the maximum *LAI* was larger. The 78.5 g C m$^{-2}$ yr$^{-1}$ annual *NEP* in 2002 was lower than those of warmer ecosystems, e.g. temperate and tropical grassland ecosystems, and as much as those of other cool ecosystems, e.g. alpine and boreal ecosystems. (2) The CO$_2$ exchange as influenced by the environmental factors was analyzed on daily and seasonal scales. The increment of photosynthetic photon flux density (*PPFD*) increased the net CO$_2$ uptake on all temporal scales. The increment of temperature decreased the *NEP* on all temporal scales, which was caused by the enhancement of ecosystem respiration in daily and seasonal changes. The increment of moisture, e.g. soil water content, atmospheric water vapor, did not affect the
daily NEP (net CO₂ uptake). The CO₂ efflux in the nighttime was suppressed by soil moisture increase. The patterns of soil water content and CO₂ efflux in the alpine ecosystem seems to be different from that reported for semi-arid grasslands.

Sim-CYCLE model simulation showed the following major findings. (3) The transition experiment for 1981-2000 showed that the annual NEP ranged from -70 to +70 g C m⁻² yr⁻¹ over the 20 years. (4) The CO₂ exchange as influenced by the environmental factors was analyzed in the interannual changes. The light increments in the growing season promoted the annual gross primary production (GPP) and NEP. Increases in temperature decreased the NEP on all temporal scales, through enhancement of the heterotrophic respiration in the interannual changes. Increases in precipitation did not affect the NEP in the interannual changes. (5) The model sensitivity analysis showed that GPP, autotrophic respiration (AR), net primary production (NPP) and plant biomass responded quickly and the heterotrophic respiration (HR), litter and soil biomass responded very slowly to the climate change. Temperature increases of 5 °C increased the GPP and those over 7.5 °C decreased the Global Warming by 5 °C advanced the ecosystem photosynthetic activity and the duration of the growing season, and increases the GPP. In contrast, warming greater than 7.5 °C may exceed the optimum temperature for photosynthesis and decreased the GPP.

This study suggests that (1) the alpine meadow was a CO₂ sink, at least in 2002. The current CO₂ sink strength is comparable with those of many sub-alpine ecosystems reported so far. The very low temperature, which suppressed photosynthetic activity and shortened the duration of the growing season, decreased the net CO₂ uptake. (2) Soil water availability was high in the alpine meadow. The high soil water might reduce the ecosystem respiration, because of the decrement of microbial activity in the well-watered anaerobic conditions. (3) The measured annual CO₂ uptake is close to the fluctuating range derived from model analysis, although CO₂ uptake data may contain estimation errors caused by energy imbalance and gap-filling methods. (4) The interannual relationships between the annual CO₂ exchange and environmental factors did not differ significantly from those of seasonal changes. (5) Long-term global warming will increase the ecosystem carbon uptake. However, extreme warming may
shift this ecosystem to another biome type through invasion of other plant species adaptive to warmer environments.

The primary conclusions are that the alpine meadow has the potential to sequester atmospheric CO₂, but the potential appears to be small, possibly because of the limitation of low temperature. However the data are not enough to judge whether this ecosystem is a CO₂ sink or source on average from the current measurements and model analysis, mainly due to the limited observation data. Multi-year data acquisitions of the CO₂ exchange are therefore required.
List of Tables

Chapter 1
Table 1.1 Global carbon budgets for the 1980s and 1990s .............................................. 7

Chapter 3
Table 3.1 Instruments and installation height ................................................................. 36
Table 3.2 Different components of carbon exchange fluxes and environmental conditions for the Kobresia humilis meadow and other sites at similar latitude .................... 37

Chapter 4
Table 4.1 Monthly meteorological conditions from August 2001 to December 2002 ...... 86
Table 4.2 Monthly heat budgets and carbon dynamics from August 2001 to December 2002 ......................................................................................................................... 87
Table 4.3 Aboveground biomass and leaf area index in 2002 ......................................... 88
Table 4.4 Parameterization of ecosystem respiration and gross primary production regression curves using 15 min averaged data ......................................................... 88
Table 4.5 Ecosystem carbon dynamics ............................................................................. 89

Chapter 5
Table 5.1 Site-specific parameters used in Sim-CYCLE running at the Haibei study site. 134
Table 5.2 Comparisons of ecosystem carbon dynamics between Sim-CYCLE estimates and those from other studies............................................................................ 135
Table 5.3 Comparisons of carbon pools and LAI between Sim-CYCLE estimates and those from other studies ..................................................................................... 137
Table 5.4 Climate and carbon fluxes annual anomalies in the Haibei alpine meadow...... 138
# List of Figures

## Chapter 2

<table>
<thead>
<tr>
<th>Figure</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>2.1</td>
<td>Location of the study area</td>
<td>10</td>
</tr>
<tr>
<td>2.2</td>
<td>Map of the observation site</td>
<td>11</td>
</tr>
</tbody>
</table>

## Chapter 3

<table>
<thead>
<tr>
<th>Figure</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>3.1</td>
<td>A schematic diagram of the flux measurement system</td>
<td>38</td>
</tr>
<tr>
<td>3.2</td>
<td>A view of the experimental site in summer</td>
<td>39</td>
</tr>
<tr>
<td>3.3</td>
<td>A view of the experimental site in winter</td>
<td>40</td>
</tr>
<tr>
<td>3.4</td>
<td>A view of the setting of the instruments - aboveground</td>
<td>41</td>
</tr>
<tr>
<td>3.5</td>
<td>A view of the setting of the instruments - belowground</td>
<td>42</td>
</tr>
<tr>
<td>3.6</td>
<td>Time series of x, y, z-axis wind velocities, air temperature, specific humidity and carbon dioxide concentration</td>
<td>43</td>
</tr>
<tr>
<td>3.7</td>
<td>Power spectrum of x, y, z-axis wind velocities, air temperature, specific humidity and carbon dioxide concentration</td>
<td>44</td>
</tr>
<tr>
<td>3.8</td>
<td>Cross spectrum between z-axis wind velocities and air temperature, specific humidity and carbon dioxide concentration</td>
<td>45</td>
</tr>
<tr>
<td>3.9</td>
<td>The wind direction dependent deflection of the 15 min average wind inclination</td>
<td>46</td>
</tr>
<tr>
<td>3.10</td>
<td>Comparisons of 15 min averaged fluxes of corrected sensible heat with uncorrected sensible heat</td>
<td>47</td>
</tr>
<tr>
<td>3.11</td>
<td>Comparisons of 15 min averaged fluxes of corrected latent heat with uncorrected latent heat</td>
<td>48</td>
</tr>
<tr>
<td>3.12</td>
<td>Comparisons of 15 min averaged fluxes of corrected CO₂ with uncorrected CO₂</td>
<td>49</td>
</tr>
<tr>
<td>3.13</td>
<td>Wind direction dependent deflection of the 15 min averaged H, LE and FCO₂ fluxes</td>
<td>50</td>
</tr>
</tbody>
</table>
Figure 3.14 Relationship between aerodynamic stability and dimensionless universal functions under unstable conditions .......................................................... 51
Figure 3.15 Relationship between aerodynamic stability and dimensionless universal functions under near neutral conditions .................................................. 52
Figure 3.16 Daily energy balance between $H+\lambda E$ and $Rn-G$ .......................................................... 53
Figure 3.17 Seasonal changes of daily energy balance ratio between $H+\lambda E$ and $Rn-G$ .......... 54
Figure 3.18 Daily changes of energy balance ratio between $H+\lambda E$ and $Rn-G$ ...................... 55
Figure 3.19 Wind direction dependent of the energy balance ratio .............................................. 56
Figure 3.20 Energy balance ratio dependent on the friction velocity ........................................... 57
Figure 3.21 Diurnal changes of hourly averaged air temperature .................................................. 58
Figure 3.22 Diurnal changes of hourly averaged soil temperature ................................................ 59
Figure 3.23 Diurnal changes of hourly averaged $PPFD$ ................................................................. 60
Figure 3.24 Mean and standard deviation of the aboveground biomass and the leaf area index for the $Koresia humilis$ alpine meadow ......................................................... 61
Figure 3.25 Diurnal changes of hourly-averaged $H$ flux for each month of the years 2001 and 2002 ............................................................................................................. 62
Figure 3.26 Diurnal changes of hourly-averaged $\lambda E$ flux for each month of the years 2001 and 2002 ............................................................................................................. 63
Figure 3.27 Diurnal courses of hourly-means of $H$, $\lambda E$ and environmental conditions in a sunny week ............................................................................................................. 64
Figure 3.28 Relationship between the hourly-means of soil surface-air temperature difference and sensible heat flux .................................................................................. 65
Figure 3.29 Relationship between hourly-means of vapor pressure deficit and latent heat flux ......................................................................................................................... 66
Figure 3.30 Diurnal changes of hourly averaged $FCO_2$ flux ......................................................... 67
Figure 3.31 Relationship between net carbon exchange rate and incident $PPFD$ in each month over a growing season ..................................................................................... 68
Figure 3.32  Diurnal courses of hourly mean CO2 exchange flux and environmental conditions ................................................................. 69
Figure 3.33  Relationship between hourly-means of the PPFD and CO2 flux ....................... 70
Figure 3.34  Diurnal courses of hourly mean gross primary production, and hourly mean ecosystem respiration ................................................................. 71
Figure 3.35  Relationship between gross primary production and irradiant PPFD, and ecosystem respiration and soil temperature in July 2002 ......................... 72
Figure 3.36  Relationship between gross primary production and incident PPFD, and GPP and evapotranspiration in 2002 ................................................................. 73

Chapter 4
Figure 4.1  Seasonal changes in meteorological conditions ............................................. 91
Figure 4.2  Seasonal changes in daily heat fluxes, moisture and canopy physiological conditions .................................................................................................................. 93
Figure 4.3  Changes in daily net ecosystem production ..................................................... 94
Figure 4.4  Linear regression of daytime accumulated CO2 uptake flux on incident photosynthetic photon flux density ................................................................. 95
Figure 4.5  Relationship between nighttime CO2 flux density and nighttime soil temperature ......................................................................................................................... 96
Figure 4.6  Relationship between nighttime CO2 flux density and nighttime soil water content during a dry spell after heavy rainfall ......................................................... 97
Figure 4.7  Ecosystem respiration calculated from light response relationships compared to the values derived from exponential regressions between soil temperature and nighttime fluxes under turbulent conditions ......................................................... 98
Figure 4.8  Seasonal changes in daily mean photosynthetic photon flux density, soil temperature, daily mean gross primary production, ecosystem respiration, ratio between GPP and Re, and daily maximum GPP and Re ............................................ 99
Figure 4.9  Changes in radiation- and water-use efficiency of gross primary production .. 100
Figure 4.10  
*LAI* controls on maximum gross primary production, radiation use efficiency and water use efficiency  ................................................................. 101

Chapter 5

Figure 5.1  Schematic diagram of Sim-CYCLE. ................................................................. 139
Figure 5.2  Schematic diagram of single leaf processes in Sim-CYCLE  ......................... 140
Figure 5.3  Schematic diagram of the Sim-CYCLE calculation processes ......................... 141
Figure 5.4  Schematic diagram of soil thermal profile processes in Sim-CYCLE ............. 142
Figure 5.5  Schematic diagram of a one-dimensional description of energy partitioning for a canopy ........................................................................... 143
Figure 5.6  Schematic diagram of runoff and infiltration processes in Sim-CYCLE .......... 144
Figure 5.7  Seasonal changes of input climate data in the steady state experiment in temperature, solar radiation, humidity and wind speed ........................................ 145
Figure 5.8  Growth of ecosystem carbon fluxes and storages, estimated by Sim-CYCLE equilibrium run ................................................................. 146
Figure 5.9  Schematic diagram of ecosystem carbon dynamics estimated by Sim-CYCLE equilibrium run ................................................................. 147
Figure 5.10 Seasonal patterns of carbon fluxes estimated by Sim-CYCLE equilibrium run 148
Figure 5.11 Seasonal patterns of carbon storages estimated by Sim-CYCLE equilibrium run ................................................................................................. 149
Figure 5.12 Seasonal patterns of soil thermal profiles, estimated by Sim-CYCLE equilibrium run ................................................................................................. 150
Figure 5.13 Seasonal patterns of heat budgets estimated by Sim-CYCLE equilibrium run.. 151
Figure 5.14 Seasonal patterns of water budgets estimated by Sim-CYCLE equilibrium run ................................................................................................. 152
Figure 5.15 Comparisons of the simulated above ground *NPP* of Sim-CYCLE transient run and observed values of other studies ........................................... 153
Figure 5.16 Yearly changes of ecosystem carbon fluxes and storages estimated by Sim-CYCLE transient run ................................................................. 154
Figure 5.17 Annual climate and anomalies in carbon fluxes ............................................. 155
Figure 5.18 Relationships between temperature and carbon fluxes anomalies .............. 156
Figure 5.19 Relationships between precipitation and carbon fluxes anomalies .......... 157
Figure 5.20 Relationships between radiation and carbon fluxes anomalies ................. 158
Figure 5.21 Anomalies in monthly climate and carbon flux ........................................... 159
Figure 5.22 Seasonal changes in the slope of regression between temperature, precipitation, short wave radiation and carbon flux anomalies ........................................ 160
Figure 5.23 Transitional changes of carbon fluxes ......................................................... 161
Figure 5.24 Transitional changes of carbon storages ..................................................... 162
Figure 5.25 Relationship between prescribed environmental changes and equilibrium carbon fluxes ............................................................................................ 163
Figure 5.26 Relationship between prescribed environmental changes and equilibrium water, heat fluxes and soil water content ........................................... 164
## List of Symbols

<table>
<thead>
<tr>
<th>Term</th>
<th>Definition (units)</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\alpha$</td>
<td>initial slope of light-GPP hyperbolic curves ($\mu$mol CO$_2$ $\mu$mol photon$^{-1}$)</td>
</tr>
<tr>
<td>ANPP</td>
<td>aboveground net primary production (Mg C ha$^{-1}$ yr$^{-1}$)</td>
</tr>
<tr>
<td>$AR$</td>
<td>autotrophic respiration (Mg C ha$^{-1}$ yr$^{-1}$)</td>
</tr>
<tr>
<td>$\beta$</td>
<td>Bowen ratio (dimensionless)</td>
</tr>
<tr>
<td>$C_p$</td>
<td>specific heat at constant pressure ($J$ kg$^{-1}$ K$^{-1}$)</td>
</tr>
<tr>
<td>$D$</td>
<td>zero plane displacement (m)</td>
</tr>
<tr>
<td>$\Delta$</td>
<td>slope of the saturation vapor pressure versus temperature curve (hPa K$^{-1}$)</td>
</tr>
<tr>
<td>$E_a$</td>
<td>activation energy ($J$ mol$^{-1}$)</td>
</tr>
<tr>
<td>$EBR$</td>
<td>energy balance ratio (dimensionless)</td>
</tr>
<tr>
<td>$EV$</td>
<td>evaporation (mm)</td>
</tr>
<tr>
<td>$FCO_2$</td>
<td>CO$_2$ flux ($\mu$mol CO$_2$ m$^{-2}$ s$^{-1}$)</td>
</tr>
<tr>
<td>$\phi$</td>
<td>universal function of all kinds of fluctuations (dimensionless)</td>
</tr>
<tr>
<td>$G$</td>
<td>soil heat flux ($W$ m$^{-2}$)</td>
</tr>
<tr>
<td>$\gamma$</td>
<td>psychrometric constant (hPa °C$^{-1}$)</td>
</tr>
<tr>
<td>$GPP$</td>
<td>gross primary production ($\mu$mol CO$_2$ m$^{-2}$ s$^{-1}$, g C m$^{-2}$ day$^{-1}$ or Mg C ha$^{-1}$ yr$^{-1}$)</td>
</tr>
<tr>
<td>$GPP_{\text{max}}, GPP_{547}$</td>
<td>maximum $GPP$ and $GPP$ at light saturation ($\mu$mol m$^{-2}$ s$^{-1}$)</td>
</tr>
<tr>
<td>$H$</td>
<td>sensible heat flux ($W$ m$^{-2}$)</td>
</tr>
<tr>
<td>$HR$</td>
<td>heterotrophic respiration (Mg C ha$^{-1}$ yr$^{-1}$)</td>
</tr>
<tr>
<td>$k$</td>
<td>von Karman’s constant (= 0.4)</td>
</tr>
<tr>
<td>$L$</td>
<td>Monin-Obukhov length (m)</td>
</tr>
<tr>
<td>$LAI$</td>
<td>leaf area index ($m^2$ m$^{-2}$)</td>
</tr>
<tr>
<td>$\lambda$</td>
<td>latent heat ($J$ kg$^{-1}$)</td>
</tr>
<tr>
<td>$\lambda E$</td>
<td>latent heat flux ($W$ m$^{-2}$)</td>
</tr>
<tr>
<td>$\lambda E_{eq}$</td>
<td>equilibrium evaporation ($W$ m$^{-2}$)</td>
</tr>
<tr>
<td>$\lambda E_{imp}$</td>
<td>imposed evapotranspiration rate ($W$ m$^{-2}$)</td>
</tr>
</tbody>
</table>
\[\text{NEE}\] net ecosystem exchange (\(\mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}\))

\[\text{NEP}\] net ecosystem production (\(\mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}, \text{ g C m}^{-2} \text{ day}^{-1} \text{ or Mg C ha}^{-1} \text{ yr}^{-1}\))

\[\text{NEP}_{\text{daytime}},\ \text{NEP}_{\text{nighttime}},\ \text{NEP}_{\text{total}}\]
net ecosystem production during daytime, nighttime and all day (\(\text{ g C m}^{-2} \text{ day}^{-1}\))

\[\text{NPP}\] net primary production (\(\mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}, \text{ g C m}^{-2} \text{ day}^{-1} \text{ or Mg C ha}^{-1} \text{ yr}^{-1}\))

\[\text{PN}\] penetration (mm)

\[\text{PPFD}\] photosynthetic photon flux density (\(\mu\text{mol photon m}^{-2} \text{ s}^{-1}, \text{ mol photon m}^{-2} \text{ day}^{-1}\))

\[\text{PR}\] precipitation (mm)

\[q'\] friction specific humidity (\(\text{g kg}^{-1}\))

\[R\] gas constant (8.134 J K\(^{-1}\) mol\(^{-1}\))

\[R_{\text{aero}}\] aerodynamic resistance (s m\(^{-1}\))

\[R_{\text{canopy}}\] canopy resistance (s m\(^{-1}\))

\[R_e\] ecosystem respiration (\(\mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}, \text{ g C m}^{-2} \text{ day}^{-1} \text{ or g C m}^{-2} \text{ yr}^{-1}\))

\[R_{e_{\text{max}}}\] ecosystem respiration (\(\mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}\))

\[R_{e_{\text{Ref}}}\] ecosystem respiration rate at \(T_{\text{ref}} (= 283.16 \text{ K}) (\mu\text{mol m}^{-2} \text{ s}^{-1})\)

\[R_{\text{a}}\] net radiation (W m\(^{-2}\))

\[RO\] runoff (mm)

\[\rho\] air density (kg m\(^{-3}\))

\[\rho_{c'}\] friction CO2 concentration (ppm)

\[\text{SWC}\] soil water content (m\(^3\) m\(^{-3}\))

\[\text{SWC}_{\text{up}}, \text{SWC}_{\text{lw}}\] soil water content (m\(^3\) m\(^{-3}\))

\[\sigma_{w}, \sigma_{T}, \sigma_{q}, \sigma_{c}\] standard deviations of vertical element of wind velocity \(w\), air temperature \(T\), specific humidity \(q\), CO2 concentration \(\rho_c\) (m s\(^{-1}\), \(^\circ\text{C}\), g kg\(^{-1}\) and ppm)

\[T_{\text{air}}, T_{\text{soil}}, T_{\text{surface}}\] air, soil and soil surface temperature (\(^\circ\text{C}\))

\[TR\] transpiration (mm)
\[
TW \quad \text{thawing snow (mm)}
\]
\[
u_* \quad \text{friction velocity (m s}^{-1}\text{)}
\]
\[
u, v, w \quad \text{x, y, z-axis wind velocities (m s}^{-1}\text{)}
\]
\[
VPD \quad \text{vapor pressure deficit (kPa)}
\]
\[
WC \quad \text{stem biomass (Mg C ha}^{-1}\text{)}
\]
\[
WF \quad \text{foliage biomass (Mg C ha}^{-1}\text{)}
\]
\[
W_{plant} \quad \text{plant biomass (Mg C ha}^{-1}\text{)}
\]
\[
WR \quad \text{root biomass (Mg C ha}^{-1}\text{)}
\]
\[
W_{soil} \quad \text{litter + mineral soil biomass (Mg C ha}^{-1}\text{)}
\]
\[
\Omega \quad \text{decoupling factor (dimensionless)}
\]
\[
Z \quad \text{reference height (m)}
\]
\[
Z_0 \quad \text{roughness length of the grassland (m)}
\]
\[
\zeta \quad \text{aerodynamic stability (dimensionless)}
\]
Chapter 1 General Introduction

1.1. Background

1.1.1. Global climate change and terrestrial ecosystems

The Earth climate system and the functioning of terrestrial ecosystem interact strongly each other. A change in one affects the other and feedback mechanisms occur. These natural interactions are now disturbed by human activities at an unprecedented scale, both in their rate and in their geographical extent. Global climate change is now in progress. Atmospheric CO₂ concentration has increased by nearly 30% since the pre-industrial 1850, when it was about 280 ppm. It is currently 370 ppm and the Earth’s surface has been warmed by 0.6 °C during the twentieth century. Climate and terrestrial ecosystems maintain their interactive relationships through CO₂, water cycles and energy flow. Therefore, climate are strongly affected by changes in the productivity of terrestrial ecosystems, and a major integrated process between living organisms and the atmosphere, and land use and feeds back on productivity of terrestrial ecosystems (Roy et al., 2001).

Considering the global CO₂ budget, there are imbalances between the CO₂ efflux by the combustion of the fossil fuel, cement production and land use change, the CO₂ uptake by the ocean known as the “missing sink” (IPCC, 1990). Eight years ago, IPCC (1996) suggested that “missing sink” may be in several forest ecosystems, tropical and boreal forests, but has not concluded. Houghton (2003) has estimated the “residual” terrestrial CO₂ flux to be as much as \(-2.9 \pm 1.1 \text{ Pg C yr}^{-1}\) in 1990s (Table 1.1). It was equivalent to one-third of the CO₂ emission by the fossil fuel and land use change (6.3 ± 0.4 and 2.2 ± 0.8 Pg C yr\(^{-1}\), respectively) and as large as the other two major CO₂ sinks; the atmosphere and the ocean (-3.2 ± 0.2 Pg C yr\(^{-1}\), -2.4 ± 0.7 Pg C yr\(^{-1}\), respectively).

The Kyoto Protocol, agreed in Kyoto COP3 Conference (the 3rd Session of the Conference of the Parties to the United Nations Framework Convention on Climate Change) held in Kyoto, Japan, December 1997, has not only set legally-binding greenhouse gas emission objectives for each industrialized country, but also allowed for an international emissions
trading scheme. During the Kyoto commitment period, carbon sinks may be included in
emissions trading by allocating credits for the amount of carbon sequestered; increase of carbon
stocks due to human induced forest establishment, afforestation and reforestation. However, the
CO₂ sequestration functioning of terrestrial ecosystems, including forests, grasslands and
deserts etc., has not been understood sufficiently, and is needed further researches using the
scientifically correct method are needed to predict the progress of global warming.

1.1.2. Measurement and modeling of carbon dioxide between the atmosphere and
terrestrial ecosystems

The "FLUXNET" research network project, including AmeriFlux, CarboEurope (former
EUROFLUX), AsiaFlux, KoFlux, OZFlux etc., will provide current understanding of how CO₂,
water cycles and energy flow and the productivities of various type ecosystems interact. The
FLUXNET is a global network of micrometeorological tower sites that use eddy covariance
methods to measure the exchanges of CO₂, water vapor, and energy between terrestrial
ecosystems and atmosphere. At present, over 200 tower sites are operating on a long-term and
continuous basis. Researchers also collect data on site vegetation, soil, hydrologic, and
meteorological characteristics at the tower sites. Valentini et al. (2000) presented data of net
ecosystem carbon exchange, collected between 1996 and 1998 from 15 EUROFLUX forest sites,
and suggested that many European forest ecosystems act as carbon sinks. Falge et al. (2002)
used the FLUXNET database to research the factors that control seasonal changes in gross
primary production (GPP) and ecosystem respiration (Rₑ) in a wide range of terrestrial
ecosystems. Their results showed that the seasonality of GPP was determined by the life-form
(e.g. broad leaf or needle leaf), and that of Rₑ was affected by the climatic types that are
characterized usually by temperature. Wilson et al. (2002) presented data of the warm season
partitioning between sensible and latent heat flux, and Bowen ratio at 27 FLUXNET sites over
66 site years. Their result showed that the climatic control on Bowen ratio was quantified using
the climatological resistance, which is proportional to the ratio of vapor pressure deficit to net
radiation, and in which there were some general differences between vegetation types and climates.

Recently, many integration models that represent carbon, water and energy exchanges between terrestrial ecosystems and the atmosphere have been developed. Based on approaches to estimate photosynthesis (GPP) or NPP, these models are divided into three categories; Light Use Efficiency (LUE) approach, Biochemical approach, and Carbon assimilation approach (Cramer et al., 1999; Arora, 2002). LUE approach, e.g. in CASA (Potter et al., 1993), BIOME2 (Haxeltine et al., 1996), GLO-PEM (Goetz et al., 2000), uses satellite data (NDVI) to determine the temporal behavior of the photosynthetic active tissue, and calculates GPP or NPP as the product of absorbed photosynthetic active radiation (APAR) by a conversion efficiency factor ε. These models can be used to examine the effect of climate variability on NPP, but the time of interest is limited to that of the satellite archive. The satellite observations provide some biosphere production. Biochemical approach, e.g. in BIOME-BGC (Running and Hunt, 1993), CARAIB (Warrant et al., 1994), DOLY (Woodward et al., 1995), BIOME3 (Haxeltine and Prentice, 1996), HYBRID (Friend et al., 1997), simulates changes in both structure (vegetation distribution and phenology) and function (biogeochemistry) of ecosystems. Generally, equilibrium between climate and vegetation is assumed, but the models can be turned into dynamical global vegetation models (DGVMs). To date, they have been applied to potential vegetation only. This is in contrast to some of models in the other categories which account for land use either explicitly (CARAIB) or implicitly through the use of satellite observations. Carbon assimilation approach, e.g. in FOREST-BGC (Running and Coughlan, 1988), TEM (McGuire et al., 1992), CENTURY (Parton et a., 1993), FBM (Ludeke et al., 1994), PnET-DAY (Aber et al., 1996), Sim-CYCLE (Ito and Oikawa, 2002), simulates the biogeochemical fluxes on the basis of soil and climate characteristics, using either vegetation maps or biogeography models to prescribe vegetation structure. These models simulate phenology either explicitly or implicitly so that the seasonal activity of a canopy can change in response to climate change.

Sim-CYCLE is a mechanistic model, on the basis of the dry-matter production theory established by Monsi and Saeki (1953). The atmosphere-biosphere CO2 exchange is composed
of physiological processes, such as gross primary production (GPP), autotrophic respiration (AR), and heterotrophic respiration (HR), and then this simulator enables us to estimate the ecosystem carbon budget in a mechanistic way. Terrestrial ecosystems were conceptualized as a five-compartment system: foliage, stem and branch, root, litter, and mineral soil. Sim-CYCLE also contains water, heat and radiation subschemes to estimate physical environment in terrestrial ecosystems. Ito and Oikawa (2000) performed a model analysis of the effect of climatic perturbations from 1970 to 1997 on the carbon budget of terrestrial ecosystems on a global scale, using Sim-CYCLE. During the 28 yr experimental period, global NEP ranged from -2.06 Pg C yr\(^{-1}\) (source) in 1983 to +2.25 (sink) Pg C yr\(^{-1}\) in 1971, being sufficiently large to give rise to anomalies in the atmospheric CO\(_2\) concentration from +0.97 to -1.06 ppmv. Regression analyses demonstrated that annual \(\Delta\text{NEP}\)s had the highest correlation with the temperature anomaly on a global scale and the responsiveness was primarily attributable to the temperature sensitivities of plant respiration and soil decomposition, and secondarily to the moisture sensitivity of decomposition. They also showed that an ENSO event and a volcanic eruption, affected global \(\Delta\text{NEP}\). Thus, the climate dependencies of global terrestrial ecosystems may contain significant implications not only for the present functioning of atmosphere-biosphere carbon exchange, but also for ongoing global warming.

### 1.1.3. An alpine meadow ecosystem on the Qinghai-Tibetan Plateau, China

On a global basis, grasslands are one of the most widespread vegetation types. Natural grasslands, including tundra, cover 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 lowland grassland ecosystems at elevations below 1500 m; studies on alpine grassland ecosystems at elevations above 3000 m are rare.
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) in spite of the restricted growing season. Additionally, low temperature in winter restricts the decomposition of litter. Therefore, this ecosystem might be an annual net sink of atmospheric CO\textsubscript{2}. Alpine meadow ecosystems in China cover approximately 63.7 \times 10^4 \text{ 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\textsuperscript{-2}; Ni, 2002) is much higher than that of savanna (5.4 kg m\textsuperscript{-2}; Adams et al., 1990) and temperate grassland (13.0 kg m\textsuperscript{-2}; Adams et al., 1990) and similar to that of tundra (22.0 kg m\textsuperscript{-2}; Adams et al., 1990), where low temperature also limits the growing season. However, tundra is reported to be at risk of changing from a sink to a source of CO\textsubscript{2} with global warming (Oechel et al., 1993). Will the alpine ecosystem on the Qinghai-Tibetan Plateau also be a net carbon source in the warming climate?

1.2. Purpose of this study

IPCC (2001) reported that if the atmospheric CO\textsubscript{2} concentration rises at a rate of 1.0% yr\textsuperscript{-1}, annual averaged temperatures in 2071-2100 on the Qinghai-Tibetan Plateau will become higher of 3.3-8.4\degree C in summer (June, July and August) and 3.2-10.9\degree C in winter (December, January and February) than those in 1961-1990. How will the plants and soil microbes react against those great warming and other climate changes? There are no series of fundamental studies on carbon dioxide flux and its dynamics to answer this question.

To fill the regional blanks in knowledge identified by global network of carbon dynamics studies, in this study, the CO\textsubscript{2} exchange between the atmosphere and an alpine meadow ecosystem was measured and simulated at the Haibei station, Chinese Academy of Sciences on the Qinghai–Tibetan Plateau, China, for short and long-term periods. First, a field measurement by the eddy covariance method was conducted to investigate the diurnal and seasonal relationships between the CO\textsubscript{2} flux and environmental factors from August 2001 to December 2002. Second, a model simulation by the Simulation Model of Carbon CYCLE in Land Ecosystem (Sim-CYCLE), developed by Ito and Oikawa (2002), has been made to investigate
the interannual relationship between the CO₂ dynamics and climate changes from 1981 to 2000 using the climate data derived at the Haibei meteorological observation field for 20 years. Finally, the differences in the relationships between the CO₂ dynamics and environmental factors on the different temporal period were analyzed and the influences of future climate changes on the CO₂ dynamics were discussed.

The objectives of this study are summarized as follows:

To clarify the daily, seasonal and interannual patterns of CO₂ flux.

To investigate the relationship between CO₂ flux and the environmental factors.

To predict the influences of global warming on the CO₂ flux and dynamics in an alpine meadow ecosystem.

1.3. Composition of the thesis

Chapter 1 introduces the relationship between the global climate change and terrestrial ecosystem, and the global network of flux measurement and model studies to investigate that relationship, and summarizes the objectives of this study. Chapter 2 describes the geographical and vegetation characteristics at the study site. Chapter 3 clarifies the diurnal changes of the CO₂ flux measured by the eddy covariance method and the relationship on the environmental factors in the field study. Chapter 4 clarifies the seasonal changes of the CO₂ flux measured by the eddy covariance method and the relationship on the environmental factors in the field study. Chapter 5 clarifies the interannual changes of the CO₂ dynamics simulated by the Sim-CYCLE and their relationships on the climate changes from 1981-2000 in the model analysis. Chapter 6 summarizes the differences in the ecosystem responses to the environmental changes between the various time scales, and predicts the influences of future climate change. Chapter 7 gives the conclusion of this study.
Table 1.1. Global carbon budgets for the 1980s and 1990s (Pg C yr\(^{-1}\)).

<table>
<thead>
<tr>
<th></th>
<th>1980s</th>
<th>1990s</th>
</tr>
</thead>
<tbody>
<tr>
<td>Fossil fuel emissions</td>
<td>5.4 ± 0.3</td>
<td>6.3 ± 0.4</td>
</tr>
<tr>
<td>Atmospheric increase</td>
<td>3.3 ± 0.1</td>
<td>3.2 ± 0.2</td>
</tr>
<tr>
<td>Oceanic uptake</td>
<td>-1.7 ± 0.6</td>
<td>-2.4 ± 0.7</td>
</tr>
<tr>
<td>Net terrestrial flux</td>
<td>-0.4 ± 0.7</td>
<td>-0.7 ± 0.8</td>
</tr>
<tr>
<td>Land-use change</td>
<td>2.0 ± 0.8</td>
<td>2.2 ± 0.8</td>
</tr>
<tr>
<td>Residual ‘terrestrial’ flux</td>
<td>-2.4 ± 1.1</td>
<td>-2.9 ± 1.1</td>
</tr>
</tbody>
</table>

Chapter 2 Site description

Measurements and estimations of carbon, water vapor and sensible heat fluxes and environmental factors were conducted at the Haibei Alpine Meadow Ecosystem Research Station, Northwest Plateau Institute of Biology, Chinese Academy of Science (lat 37°29′–45°N, long 101°12′–23°E; 3250 m a.s.l.; Figs. 2.1 and 2.2). The research station is located in a large valley oriented northwest–southeast in northeast Qinghai-Tibetan Plateau, and surrounded on all sides by the Qilian Mountains. The average altitude of the mountains and the valley are 4000 and 2900–3500 m a.s.l. The Datong River passes from north 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 from May to September, the plentiful sunshine and rainfall (80% of annual total rainfall) allow plants to grow effectively, although harmful UV-B radiation is high (Li and Zhou, 1998).

The soil is a clay loam; its average thickness is 65 cm. The surface 5–10 cm horizons, which are classified as Mat Cry-gelic Cambisols according to the Chinese national soil survey classification system (Institute of Soil Science and Chinese Academy of Sciences, 2001), are wet and rich in organic matter.

The plant community is dominated mainly by three major 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 (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 (Li and Zhou, 1998). During the growing season, the plants accumulate photosynthates in belowground storage organs for new shoot
growth in next spring. This characteristic is a feature of alpine grassland plants (Li and Zhou, 1998).
Figure 2.1. Location of the study area
Figure 2.2. Map of the observation site

- Flux measurement site
- Haibei station, CAS
Chapter 3 Diurnal changes of the CO2 exchanges

3.1. Introduction

The Qinghai-Tibetan Plateau covers vast area, about $2.5 \times 10^6$ km$^2$ (Zheng et al., 2000), and may contribute significantly to the CO$_2$, water and energy exchange between the atmosphere and vegetation on a regional or even a global scale. However, in contrast with extensive information obtained in recent years about water and energy exchange (Yasunari, 2001), little evidence is available for understanding CO$_2$ exchange and its relation to environmental controls in this unique ecosystem.

Eddy covariance method that acquires the fluctuations of wind speed, CO$_2$ concentration etc. directly, enable us to measure the CO$_2$ flux without any assumption and has been used widely in various terrestrial ecosystems (Aubinet et al., 2000; Baldocchi et al., 2001; Yamamoto et al., 2001). Net ecosystem exchange (NEE) data measured by this method shows that knowledge of leaf and whole plant physiology can be used to interpret whole system variability (e.g. Hollinger et al., 1994; Amthor et al., 1994). Furthermore, Valentini et al. (2000) divided daytime NEE data from a global network of tower observations of flux into gross primary production (GPP) and ecosystem respiration ($R_e$) by extrapolating site-specific exponential relationships between nocturnal soil temperature and CO$_2$ efflux. This approach allows the diurnal pattern and amplitude of GPP and $R_e$ to be investigated.

From August 2001 to December 2002, long-term CO$_2$, water vapor and energy flux measurements using the eddy covariance method were carried out in an alpine meadow on the Qinghai-Tibetan Plateau. Additionally, measured net ecosystem exchanges were divided into GPP and $R_e$. The aims of this chapter are 1) to show the diurnal changes NEE, GPP and $R_e$, 2) to clarify the environmental controls on the diurnal changes of the CO$_2$ fluxes.

3.2. Materials and Methods
3.2.1. Micrometeorology

Micrometeorological measurements were conducted at the winter pasture in the Haibei station of CAS from 9 August to 31 December 2002 (Fig. 2.2). The study site is flat with a fetch of at least 250 m in all directions. 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; Table 3.1) and a PPFD meter (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 points with heat plates (HFT-3, Campbell Scientific Inc., Logan, UT, USA). 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 manufactured thermocouples. Soil water content was measured at 0.05, 0.2, and 0.5 cm 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.) (Fig. 3.1-3.5).

3.2.2. Eddy covariance method

3.2.2.1. Fundamentals of the eddy diffusion flux calculations

1. Definition of the turbulent flow

Figure 3.6 shows the time series of x, y, z-axis wind velocities (u, v, w), air temperature (\(T_a\)), specific humidity (q) and carbon dioxide concentration (\(\rho_c\)) at the Haibei grassland in winter, 2003 at the frequency of 20 Hz by a sonic anemometer and an open-path Infra-red gas analyzer (the details are in the section 3.2.2.2). Those scalar values were disturbed, and there
seems to be no relationship with each other. This fluctuation is caused by the eddy generation and passage, called as a ‘turbulent flow’.

2. Calculation of the turbulent transfer

The vertical turbulent transfer of substance and energy is generated by the turbulent flow and the vertical gradient of substance concentrations and temperatures. The turbulent flux of voluntary substance or energy is represented as follows:

\[ Q_c = \overline{w} \rho C \]  \hspace{1cm} (3.1)

\[ = \rho \overline{wC} \]  \hspace{1cm} (3.2)

where \( Q_c \) is the turbulent flux of substance or energy (kg m\(^{-2}\) s\(^{-1}\)), \( w \) is the vertical transfer velocity of air (m s\(^{-1}\)), \( C \) is the voluntary substance concentration (kg kg air\(^{-1}\)), \( \rho \) is the air density (kg m\(^{-3}\)). The upper bar means an average. The air density is assumed to be constant in Eq. (3.2). \( w \) and \( C \) are divided into average and fluctuation components as follows;

\[ w = \overline{w} + w' \]  \hspace{1cm} (3.3)

\[ C = \overline{C} + C' \]  \hspace{1cm} (3.4)

where the prime mark means fluctuation value. Eqs. (3.3) and (3.4) are inserted into Eq. (3.2) as follows:

\[ Q_c = \rho \overline{wC} + \rho \overline{w'C} + \rho \overline{wC'} + \rho \overline{w'C'} \]  \hspace{1cm} (3.5)

\[ = \rho \overline{w'C'} \]  \hspace{1cm} (3.6)

where \( \overline{w'} \) and \( \overline{C'} \) are zero. \( \overline{w} \), the average vertical component of air transfer, is assumed to be zero. When sensible heat \( (H) \), latent heat \( (\lambda E) \) and carbon dioxide fluxes \( (FCO_2) \) are
calculated, \( C \) is assigned for air temperature \( T = T' + T'' \) (K), a specific humidity \( q = q + q' \) (kg kg\(^{-1}\)) and atmospheric CO\(_2\) concentration \( \rho_c = \rho_c + \rho_{c}' \) (mg CO\(_2\) m\(^{-3}\)), respectively as follows:

\[
H = C_p \rho w'T'
\]

(3.7)

\[
\lambda E = \lambda \rho w'q'
\]

(3.8)

\[
FCO_2 = w' \rho_c'
\]

(3.9)

where \( \rho \) is the air density (kg m\(^{-3}\)), \( C_p \) is the specific heat at constant pressure (J kg\(^{-1}\) K\(^{-1}\)), \( \lambda \) is the latent heat (J kg\(^{-1}\)) as shown in the next:

\[
\rho = 100 \times P / \left( 287.1 \times \left( T_a + 273.16 \right) \times \left( 1 + 0.61 \times q \right) \right)
\]

(3.10)

\[
C_p = 1004.7
\]

(3.11)

\[
\lambda = \left( 2.501 - 0.00237 \times T_s \right) \times 10^6
\]

(3.12)

where \( P \) is the air pressure (hPa), \( T_a \) is the air temperature (°C), and \( T_s \) is the soil surface temperature (°C).

3. Power spectrum of turbulent fluctuation - Verification of turbulent measurement -

To confirm the successful acquisition of turbulent flow across all frequency bands contributing the flux calculation, the power spectrums of each scalar fluctuation should be checked. If acquiring successfully, the spectrum in high frequency band should conform the concept of an inertial sub-range, a part of equilibrium range of turbulent flow (Kaimal, 1988). An equilibrium range is high frequency zone that comprises the local isotropic under which condition the low frequency eddy, caused by the obstruction, is transferred by inertial force and
pressure. The inertial sub-range exist in the frequency range, in which energy does not generate and disappear and energy is only handed off to smaller scale filed, between the energy holding frequency \( k_0 \) and the viscosity dispersing frequency \( k_d \). This sub-range usually generated in the atmospheric boundary layer in the condition of high Reynolds number. In this sub-range, the energy spectrum \( F_x(\kappa) \), is with no statistically relation to energy holding eddy and viscosity relevant eddy, is determined by energy dissipation rate \( \varepsilon \) exclusively and described in the dimension analysis as follows:

\[
F_x(\kappa) = \alpha \varepsilon^{2/3} \kappa^{-3/5}
\]  

where \( \kappa \) is wave number and \( \alpha \) is defined as an absolute constant.

Existence of inertial sub-range in the energy spectrum shows that the eddy covariance measurement instruments acquire the turbulent flow through high frequency zone normally. Although the spectrum theory is usually formulated in wave number space, in the most of turbulent flow measurement, the spectrum is able to be converted into the frequency scale in their spatial scales using the formulation of a Taylor’s hypothesis. In the \( u \)-spectrum, conversion is described as follows:

\[
\kappa = 2\pi f \sqrt{u}
\]

where \( f \) is the frequency. The spectrum of frequency \( Su(f) \) is defined as follows:

\[
\int_0^{\infty} F_x(\kappa) \, d\kappa = \sigma_s = \int_0^{\infty} Su(f) \, df
\]

where \( \sigma_s \) is variances of wind speed and substance concentration. Eq. (3.15) is assigned by Eq. (3.14) and differentiated by \( f \) as follows:
\[
\frac{2\pi}{u} Fx \left( \frac{2\pi}{u} \right) = Su(f) \quad (3.16)
\]

that is:

\[
\kappa Fx(k) = fSu(f) \quad (3.17)
\]

Next, the similarity function \( \phi_e \) of the dissipation rate of turbulent movement energy \( E \) is defined as follows:

\[
\phi_e = k\varepsilon \gamma / u^3 \quad (3.18)
\]

where \( k \) is the karman constant (dimensionless), \( z \) is height (m), \( u_* \) is the friction velocity (m \( s^{-1} \)).

The \( u \)-spectrum of inertial sub-range is presented in conformity with similarity theory of surface boundary layer as follows:

\[
\frac{fS_u(f)}{u^2} = \frac{\alpha_1}{(2\pi)^{2/3}} \left( \frac{\varepsilon^{2/3} z^{2/3}}{u^2} \right) \left( \frac{f}{u} \right)^{-2/3}
\]

\[
= \frac{\alpha_1}{(2\pi k)^{2/3}} \phi_e^{2/3} \left( \frac{f}{u} \right)^{-2/3} \quad (3.19)
\]

where \( \alpha_1 \) is uniform value. If the \( f \gamma / u \) is assumed as a dimensionless frequency \( n \):

\[
\frac{fS_u(f)}{u^2 \phi_e^{2/3}} = \frac{\alpha_1}{(2\pi k)^{2/3}} n^{-2/3} \quad (3.20)
\]
The spectrum in the inertial sub-range is described by a line with the slope of $-2/3$ in the logarithmic graph.

Figure 3.7 shows the power spectrum of $x, y, z$-axis wind velocities ($u, v, w$), air temperature ($T$), specific humidity ($q$) and carbon dioxide concentration ($C$) measured at the Haibei grassland for 12:00-12:15, 13, Aug., 2002. Vertical axes of the figures are normalized with the frequency. The spectrums of each variable are proportional to frequencies in the slope of $-2/3$ within the inertial sub-range. This confirms that the measurements of turbulent fluctuations are acquired normally in this study.

4. Cross spectrum of turbulent fluctuation - Determination of sampling frequency and calculating period-

All turbulent fluctuation, which contributes to flux calculations, should be acquired between sampling (cut-off) frequency and average time periods, and calculated flux must retain constancy. These frequency and time periods should be determined after checking the cross-spectrum of each fluctuation.

Wyngaard and Cote (1972) present that the cross-spectrum declines in the slope of $-7/3$ within the inertial sub-range of $uw$ and $wT$ in their empirical equation. In this sub-range which $z/L$ and $n$ are only normalized cross-spectrum, $w$ and $T$ are described in the logarithmic scale as follows:

$$-\frac{fC_{wT}(f)}{u^2} \propto G(z/L)n^{-4/3}$$

where $G(z/L)$ is determined experimentally as a function of $z/L$.

Figure 3.8 shows the cross spectrum between $z$-axis wind velocities ($w$) and air temperature ($T$), specific humidity ($q$) and carbon dioxide concentration ($C$) at the Haibei grassland for 12:00-12:15, 13, Aug., 2002. Vertical axes of the figure are normalized with the frequency. The cross spectrums of each variable are proportional to frequencies in the slope of
-4/3 within the inertial subrange. The cross spectrum between vertical wind $w'$ and air temperature, specific humidity and CO2 concentration $c'$ showed that co-variance was close to zero when the noise was at frequencies lower than 0.002 Hz (8.3 min per cycle; Fig. 3.8). A period of 15 min was thus enough to avoid the effect of noise. The mean, variance, and covariance values were then calculated and logged for every 15 min.

3.2.2.2. Measurement of eddy diffusion fluxes

Wind speed and sonic virtual temperature were measured at 2.2 m above the ground with a sonic anemometer (CSAT-3, Campbell Scientific Inc.). Carbon dioxide and water vapor concentrations were also measured at the same height with an open-path infra-red gas analyzer (CS-7500, Campbell Scientific Inc.). Wind speed, sonic virtual temperature, and CO2 and H2O concentrations were sampled by the digital micrologger at a rate of 10 Hz.

3.2.2.3. Correction of measured turbulent flux data

1. WPL correction

In a CO2 flux calculation, the air density variation was taken into account in the calculation processing of turbulent flux, as supposed by Webb et al. (1980). CO2 flux ($F_{CO2}$) was given in the next:

$$ F_{CO2} = \overline{w'c'} + \overline{w'\rho_q} \left( \mu \frac{\overline{c'}}{\rho_a} + \frac{\rho_a}{\rho_a} \right) + \frac{\overline{c'}}{\sigma} \left( \frac{\overline{w'T'}}{T} \right) $$

(3.22)

where $F_{CO2}$ is the CO2 flux (mg CO2 m$^{-2}$ s$^{-1}$), $\rho_q$ and $\rho_a$ are densities of water vapor and air (kg m$^{-3}$), $\mu$ is the ratio of dry air mass to water vapor mass (kg kg$^{-1}$), $\sigma$ is the ratio of $\rho_q$ to $\rho_a$.

2. Coordinate rotation, trend removing, and water vapor correlation

In this study, flux data correction, e.g. coordinate rotation, trend removing and water vapor correlation, are not enforced to all sampled fluctuation data. However, the influences of un-correction on the calculated flux are examined for 10 days in July, 2002, using fluctuation
data sampled at the frequency of 10 Hz, and implicit estimation error in the flux data are discussed.

Coordinate Rotation Although sonic anemometer is usually settled horizontally, slight inclination of instruments will generate for long-term experiment. The landform, installation mast and instrument body also generate the wind disturbance and bias the horizontal wind (Fig. 3.9) and $\bar{w}$ will not become zero. In that situation, it is necessary to rotate the coordinates of wind vectors by the degrees of blowing up and down angle ($\psi$).

First, horizontal wind directions are rotated as follows:

$$\theta = \tan^{-1} \left( \frac{U_y}{U_x} \right)$$

$$U = U_x \cos \theta + U_y \sin \theta$$  \hspace{1cm} (3.23)

$$V = -U_x \sin \theta + U_y \cos \theta$$  \hspace{1cm} (3.24)

where $\theta$ is the degrees of anticlockwise rotation angle from the center line of anemometer, $U_x$ and $U_y$ are horizontal and vertical wind velocities in the direction toward the center line of anemometer, $U$ and $V$ are x- and y-axis wind velocities after coordinate rotation. After this coordinate rotation, $\bar{U}$ and $\bar{V}$ become zero.

Second, vertical wind directions are rotated as follows:

$$\psi = \tan^{-1} \left( \frac{W}{U} \right)$$

$$U_e = U \cos \psi + W \sin \psi$$  \hspace{1cm} (3.26)

$$W_e = -U \sin \psi + W \cos \psi$$  \hspace{1cm} (3.27)
where $\psi$ are the degrees of blowing up and down angle, $\overline{U}$ and $\overline{W}$ are average horizontal and vertical wind velocities, $U_c$ and $W_c$ are $x$- and $z$-axis wind velocities after coordinate rotation. After this coordinate rotation, $U_c$ and $W_c$ become zero.

Trend removing  Gradual changes of air temperature and humidity in the morning and evening, or instrumental signal drift may bias the averaged data during 15 min average period. In those cases, the differences from the average will be bigger apparently, and this affects the turbulent statistical values, e.g. standard deviation and covariance. Consequently, trend values derived from regression line against time series, are subtracted from the fluctuation data.

Water vapor correction  Virtual sonic temperatures derived from sonic anemometer, was affected from the fluctuations of atmospheric water vapor. The water vapor correction is necessary to measure temperature fluctuations and calculate sensible heat flux accurately (Tsukamoto et al., 2001).

\[
T = T_o (1 - 0.514q)
\]

(3.28)

where $T$ is corrected air temperature ($^\circ$C), and $T_o$ is uncorrected air temperature ($^\circ$C).

Figures 3.10, 3.11, 3.12 make comparisons between corrected fluxes and uncorrected fluxes in sensible and latent heat and CO₂ flux calculations. Each regression line slope shows small differences, within 4%, between corrected fluxes and uncorrected fluxes. This indicates that it is unnecessary to apply these three corrections to turbulent fluctuation data.

3.2.2.4. Validity of measured flux data

1. Flux dependent on the wind direction

Figure 3.13 shows the dependence of sensible and latent heat and CO₂ fluxes on the wind direction. The black line in the figure indicates moving average of 100 points. Sensible heat flux $H$ fluctuates between 90 and 270 degrees, and show two minimum peaks in 160 and
260 degrees. This holds in the case of latent heat $\lambda E$ and also hold in the case of CO$_2$ flux $FCO_2$ in inverse. It is considered to be due to biased average wind flow, as shown in Figure 3.8, disturbed by installation mast and data logger box settled in the direction of 180 degrees. However, moving average line shows small bias in the measured fluxes, indicating that it is unnecessary to rotate coordination of turbulent fluctuation data.

2. Dimensionless universal function

(1) Monin-Obukhov similarity

According to Monin and Obukhov (1954), in ideal surface boundary layer, i.e. keeping steady state and horizontal homogeneity, all kinds of statistics concerning wind velocity and temperature in the turbulent flow, can be described by momentum $\tau/\rho$, vertical heat flux $H/C_p\rho$ and buoyancy parameter $g/T$ at ground surface, except for height $z$. This is called “Monin-Obukhov similarity theory”.

From the three fundamental quantities mentioned above, dimensional scales of velocity, temperature and length, which describe the statistics of wind velocity and temperature, are drawn as follows:

\[ u_* = \left( \frac{\tau}{\rho} \right)^{1/2} = \left( -\overline{u'w'} \right)^{1/2} \quad (3.29) \]

\[ T_* = \frac{H}{C_p\rho u_*} = -\frac{\overline{w'T'}}{u_*} \quad (3.30) \]

\[ L = \frac{-u_*^3T_0}{kgw'T_0} \quad (3.31) \]

where $\tau$ is momentum (kg m$^{-1}$ s$^{-2}$), $T_*$ is friction temperature (K), $L$ is Monin-Obukhov length (m), $g$ is acceleration of gravity ($= 9.8$ m s$^{-2}$), and $T_0$ is temperature near ground surface ($^o$C)

According to Monin-Obukhov similarity theory, the statistics of wind velocity and temperature are described by $z/L$ exclusively as follows:
\[
\frac{F}{F_*} = g_F \left( \frac{z}{L} \right)
\]  
(3.32)

where \( F \) is the statistics of wind velocity and air temperature near ground surface, \( F_* \) is the scales of velocity, temperature and length, and \( g_F \) is universal function of \( F \).

(2) Standard deviation of turbulent fluxes \( \sigma \) and dimensionless universal function \( \phi \)

According to Monin-Obukhov similarity theory, the standard deviations of wind velocity, air temperature, specific humidity and \( CO_2 \) concentration are expressed by \( z/L \) exclusively as follows:

\[
\frac{\sigma_w}{u_*} = \phi_w \left( \frac{z}{L} \right) \quad (3.33)
\]

\[
\frac{\sigma_T}{T_*} = \phi_T \left( \frac{z}{L} \right) \quad (3.34)
\]

\[
\frac{\sigma_q}{q_*} = \phi_q \left( \frac{z}{L} \right) \quad (3.35)
\]

\[
\frac{\sigma_{\rho_c}}{\rho_{c*}} = \phi_{\rho_c} \left( \frac{z}{L} \right) \quad (3.36)
\]

where \( \sigma_w \), \( \sigma_T \), \( \sigma_q \) and \( \sigma_{\rho_c} \) are standard deviations of vertical element of wind velocity \( w \), air temperature \( T \), specific humidity \( q \), \( CO_2 \) concentration \( \rho_c \), \( q_* \) is friction specific humidity \((=E/\lambda_{wu} = \overline{w'q'}/u_*\)), \( \rho_{c*} \) is friction \( CO_2 \) concentration \((=FCO_2/u_* = \overline{w'\rho_c'}/u_*\)) and \( \phi \) is universal function of all kinds of fluctuations.

(3) \( \phi \) under strongly unstable condition
Strongly unstable condition, under which vertical convection becomes near natural convection state, makes \( L \) extremely small; \( H \) becomes bigger and \( u_* \) becomes near zero. In the case, \( L \) is useless scale and universal functions are described by \( L \), which eliminates \( u_* \) in itself, as follows:

\[
\phi_w = \frac{\sigma_w}{u_*} \approx \left( -\frac{z}{L} \right)^{\frac{1}{3}}
\]

(3.37)

\[
\phi_T = \frac{\sigma_T}{T_*} \approx \left( -\frac{z}{L} \right)^{-\frac{1}{3}}
\]

(3.38)

\[
\phi_q = \frac{\sigma_q}{q_*} \approx \left( -\frac{z}{L} \right)^{\frac{1}{3}}
\]

(3.39)

\[
\phi_{\rho_c} = \frac{\sigma_{\rho_c}}{\rho_{c*}} \approx \left( -\frac{z}{L} \right)^{\frac{1}{3}}
\]

(3.40)

Thus, the dimensionless universal functions are in relation to cube root or minus cube root of stability under strongly unstable condition. These are founded in the field experiments, e.g. Kader and Yaglom (1990), Kaimal and Finnigan (1994) and Ohtaki (1985).

Figure 3.14 shows the relationship between aerodynamic stability and dimensionless universal functions under unstable conditions from 14 Jan. to 31 Dec., 2002. Data show 15 min average values. Universal functions are distributed along the lines with the slopes of \( 1/3 \) (\( \sigma_w \)) or \(-1/3 \) (\( \sigma_T, \sigma_q, \sigma_{\rho_c} \)) against the stability \( -\zeta \) under unstable condition (\( -\zeta >= 0.2 \)), and this confirms the reasonability of sampling turbulent fluxes in this study.

(4) \( \phi \) under near neutral condition

Figure 3.15 shows the relationship between aerodynamic stability and dimensionless universal functions under near neutral conditions from 14 Jan. to 31 Dec., 2002. Data shows 15 min average values. The geometric means of \( \sigma_w/u_* \) for neutral condition (\( -0.1 < z/L <0.1 \)) was 1.08 in this study site and is similar to 0.93-1.14 reported by Yaglom (1977) and slightly lower.
then 1.2 reported by Ohtaki (1985) for the paddy field, 1.2-1.4 compiled by Pasquille (1974). The $\sigma_f/T_s$, $\sigma_f/q_s$, and $\sigma_v/\rho c_w$ are scattered for neutral condition ($-0.1 < z/L < 0.1$), but and are also distributed around the reported mean values; $\sigma_f/T_s$, 2.4 reported by Phelps and Pond (1970) for over water, $\alpha_f/q_s$, 2.6 reported by Ohtaki (1985) and 2.3 reported by Phelps and Pond (1970) for over water, $\sigma_v/\rho c_w$, 3.5 reported by Ohtaki (1985) as a geometric mean ($-z/L$ 0-0.14).

3. Energy imbalance

Historically, energy balance closure has been accepted as an important test of eddy covariance data, and a number of individual sites within the FLUXNET network report energy balance closure as a standard procedure (Wilson, et al, 2002). In this study, the energy balance ratio (EBR) between $H+\lambda E$ and $Rn-G$ is also examined.

Figure 3.16 shows the energy balance during 2001 and 2002. The regression coefficients show the surface energy fluxes ($H+\lambda E$) are underestimated by 41 % (2001) and 30 % (2002) relative to estimates of available energy ($Rn-G$). The seasonal changes of EBR may be constant for a whole year (Fig. 3.17), but the EBR is slightly larger in summer than in winter. The EBR shows the apparent diurnal changes in Figure 3.18. During morning and evening transition periods, when the mean value of $Rn-G$ was close to zero, the EBR is not especially meaningful. The EBR dependence on the wind direction, shown in Fig. 3.19, was lower in near 160° than other directions. The effect of friction velocity on the EBR was analyzed separately for daytime and nighttime data using all 15 min data in 2002 (Fig. 3.20). In daytime, when the friction velocity exceeded 0.3 m s$^{-1}$, the EBR might be convergent to near 0.7. In nighttime, there was no obvious trend in the relationship.

3.2.3. Omega factor calculation

The Penman-Montieth model, the most popular evapotranspiration model, uses an electrical analogy to treat vertical water vapor movement in the Soil-Plant-Air-Continuum (SPAC) circulation, and is the one-dimensional model of crop evapotranspiration from a single source. Evapotranspiration from the canopy to atmosphere is obtained as follows:
\[
\lambda E = \frac{\Delta (R_a - G) + \rho c_p D / r_a}{\Delta + \gamma (1 + r_c / r_a)}
\] (3.41)

in which \( \Delta \) is the slope of the saturation vapor pressure versus temperature curve (hPa K\(^{-1}\)), \( \rho \) is the density of air (kg m\(^{-3}\)), \( c_p \) is the specific heat at constant pressure (J kg\(^{-1}\) K\(^{-1}\)), \( D \) is the vapor pressure deficit (h Pa), \( r_c \) is the canopy resistance (s m\(^{-1}\)), and \( r_a \) is the aerodynamic resistance (s m\(^{-1}\)). Jarvis and McNaughton (1986) rewrote the Penman-Monteith model as follows:

\[
\lambda E = \Omega \lambda E_{eq} + (1 - \Omega) \lambda E_{imp}
\] (3.42)

\[
\lambda E_{imp} = \frac{\rho c_p D}{\gamma r_c}
\] (3.43)

\[
\Omega = \frac{\Delta + \gamma}{\Delta + \gamma (1 + r_c / r_a)} \quad (0 \leq \Omega \leq 1)
\] (3.44)

where \( \lambda E_{imp} \) is the imposed evapotranspiration rate and \( \Omega \) is the decoupling factor. When \( \Omega \) is near 1, the evapotranspiration rate is almost dominated by the available energy. \( \lambda E_{eq} \), the 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 found 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 (Jones, 1992):

\[
\lambda E_{eq} = \frac{\Delta (R_a - G)}{\Delta + \gamma}
\] (3.45)
where $lE_{eq}$ is the equilibrium evaporation (W m$^{-2}$), $\Delta$ is the slope of the relation between saturation vapor pressure and temperature (hPa °C$^{-1}$), $R_n$ is the net radiation (W m$^{-2}$), $G$ is the soil heat flux (W m$^{-2}$), and $\gamma$ is the psychrometric constant (hPa °C$^{-1}$).

### 3.2.4. Bowen ratio calculation

The Bowen ratio $\beta$ is defined as follows:

$$\beta = H/lE$$

(3.46)

### 3.2.5. Canopy resistance calculation

Canopy resistance, or diffusion flux resistance, (i.e. the stomatal resistance of a “big leaf”) is determined by the physiological activity of the canopy plants in the Penman–Monteith evaporation theory (Monteith, 1972). By transformation of the Penman–Monteith equation, canopy resistance is derived from environmental factors as follows:

$$R_{canopy} = R_{aero} \left[ \frac{\Delta (R_n - G - lE) + \rho c_p VPD/R_{aero}}{\gamma lE} - 1 \right]$$

(3.47)

where $R_{canopy}$ is the canopy resistance (s m$^{-1}$); $R_{aero}$ is the aerodynamic resistance ($=1/ku \ln(Z-d/Z_0)$ (s m$^{-1}$)); $k$ is von Karman’s constant ($= 0.4$); $u_*$ is the friction velocity (m s$^{-1}$); $Z$ is the reference height ($= 2.2$ m); $d$ is the zero plane displacement (m) ($= 0.63 \times$ canopy height (m)); $Z_o$ is the roughness length of the grassland (m) ($= 0.13 \times$ canopy height (m); Montieth, 1973).

### 3.2.6. GPP and $R_e$ calculations

Ecosystem respiration $R_e$ was measured directly during nighttime periods with strong turbulence (as a $NEE$ at a friction velocity $u_* > 0.2$ m s$^{-1}$), and was extrapolated to other periods by using exponential regressions of measured $R_e$ with soil temperature at a depth of 5 cm with an Arrhenius equation reported by Lloyd and Taylor (1994):
\[
R_e = R_{e, \text{ref}} \exp \left[\frac{E_a}{R} \left(\frac{1}{T_{\text{ref}}} - \frac{1}{T_K}\right)\right],
\]

where \( R_e \) is nighttime ecosystem respiration rate (\( \mu \text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1} \)), \( R_{e, \text{ref}} \) is ecosystem respiration rate at \( T_{\text{ref}} \) (\( = 283.16 \text{ K} \)), and \( E_a \) is the activation energy (J mol\(^{-1}\)). These two parameters are assigned as the site-specific parameters. \( R \) is a gas constant (8.134 J K\(^{-1}\) mol\(^{-1}\)), and \( T_K \) is the soil temperature at a depth of 5 cm \( T_{\text{soil}} \). \( R_{e, \text{ref}} \) was evaluated for every month. \( E_a \) was evaluated from a regression of all \( R_e \) data in 2002 against \( T_{\text{soil}} \) as a constant value throughout the year (81 519 J mol\(^{-1}\)).

Values of \( GPP \) was calculated as the difference between \( NEP \), as a negative value of \( NEE \), and \( R_e \) as follows:

\[
GPP = NEP + R_e,
\]

where \( GPP \) is ecosystem gross primary production (\( \mu \text{mol m}^{-2} \text{ s}^{-1} \)), and \( NEP \) is net ecosystem production as \( \text{CO}_2 \) uptake flux (\( -NEE \)) (\( \mu \text{mol m}^{-2} \text{ s}^{-1} \)).

### 3.2.7. LAI and biomass sampling

Aboveground biomass and \( LAI \) were investigated within a radius of 250 m around the measuring station once in 2001 (20 August) and nine times from May to September in 2002. The aboveground plants were cut at ground levels in five randomly placed 0.5-m\(^2\) 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.

### 3.3. Results

#### 3.3.1. Micrometeorology and vegetation growth
Figure 3.21 shows the seasonal changes of monthly averaged diurnal changes of air temperature $T_{air}$. In summer, July 2002, $T_{air}$ reached a minimum of 5.5 °C at 0600 and a maximum of 16.9 °C at 1500. In winter, January 2002, $T_{air}$ reached a minimum of −24.8 °C at 0900 and a maximum of −8.1 °C at 1700. Thus, this alpine meadow is characterized by a large diurnal amplitude of air temperature variation and extremely low temperature in winter. Figure 3.22 shows the diurnal changes of monthly averaged soil temperature $T_{soil}$ at the depth of −5 cm. In summer, July 2002, $T_{soil}$ reached a minimum of 14.0 °C at 0800 and a maximum of 20.7 °C at 1700. In winter, January 2002, $T_{soil}$ reached a minimum of −6.2 °C at 0900 and a maximum of −3.4 °C at 1900. Figure 3.23 shows the seasonal changes of monthly averaged diurnal changes of photosynthetic photon flux density (PPFD). In summer, May 2002, PPFD reached a maximum of 1660.9 μmol photon m$^{-2}$ s$^{-1}$ at 1200. Thus, this alpine meadow is also characterized by a large amplitude of light intensity.

In 2001, $LAI$ and aboveground biomass were 3.1 and 347 g d. w. m$^{-2}$, respectively, on 20 August (DOY233), which were the annual maximum values. In 2002, the values increased from late May (DOY145); $LAI$ reached a maximum of 3.8 on 16 July (DOY197) and then decreased slowly (Fig. 3.24). The aboveground biomass reached a maximum of 283 g d. w. m$^{-2}$ on 30 July (DOY211), remained high during August, and then decreased rapidly in September.

### 3.3.2. Sensible and latent heat fluxes

To investigate the diurnal courses and amplitudes of sensible $H$ and latent heat fluxes $\lambda E$, monthly averaged diurnal courses are shown in Figs. 3.25 and 3.26. The amplitudes of $H$ were large in April when plants were beginning to grow and in October when they were dead, but small in January in winter and July when aboveground biomass peaked. On the other hand, the amplitudes of $\lambda E$ were large in July and August when aboveground biomass peaked, but so small from October to March.

To investigate the relationship between these heat fluxes and environmental factors, the diurnal courses of them during a sunny week, 9-15 August 2001 (DOY221-227), were shown in Fig. 3.27. $H$ paralleled the differences between $T_{air}$ and $T_{surface}$, and their relationship was
apparent linear (Fig. 3.28). $\lambda E$ seems to parallel available radiation ($Rn-G$), however the afternoon $\lambda E$ was bigger than the forenoon $\lambda E$ (Fig. 3.29). The decoupling factor $\Omega$, introduced by Jarvis and MacNaughton (1986), kept nearby constant value during 9:00-16:00 and decreased after 17:00. This presents that high VPD lead to high $\lambda E$ in the evening when $Rn-G$, drove $\lambda E$ in the afternoon, decreased. The Bowen ratio, the ratio of $H$ to $\lambda E$, decreased gradually in the afternoon. This indicates that in the afternoon $\lambda E$ kept high value and, $H$ decreased with the decreasing $T_{air} - T_{surface}$.

3.3.3. Carbon dioxide flux

The author examined the seasonality of the $FCO_2$ diurnal course (Fig. 3.30). The minimum $FCO_2$, i.e. maximum uptake, was $-10.8 \mu$mol m$^{-2}$ s$^{-1}$ at 1300 in August 2001. The maximum $FCO_2$, i.e. maximum release, was 4.4 $\mu$mol m$^{-2}$ s$^{-1}$ at 0100 in August 2002. The amplitudes of $FCO_2$ were large in July and August when aboveground biomass peaked. $FCO_2$ was small in May and June when plants were beginning to grow, and in September when they were dead. In winter, October to April, the $FCO_2$ was positive even in the daytime. The time courses of $FCO_2$ were similar for August 2001 and July 2002, but the amplitude of $FCO_2$ was larger in August 2001 than in August 2002.

$FCO_2$ is plotted against PPFD in Fig. 3.31. $FCO_2$ decreased as PPFD increased in any month. During the period from July to August, however, the afternoon $FCO_2$ was higher than the forenoon $FCO_2$, and $FCO_2$ values in June, July, and August were smaller than those in May and September. Light intensity was similar between August 2001 and August 2002, but the magnitude of $FCO_2$ was larger in August 2001.

During 9-15 August 2001 (DOY221-227), a sunny week, $FCO_2$ was positive (i.e. efflux) for several hours after 0000 (midnight, Beijing Standard Time), and then became negative (i.e. uptake flux) at 0800 (Fig. 3.32). $FCO_2$ decreased to a minimum of $-12.4 \mu$mol m$^{-2}$ s$^{-1}$ at 1300 as PPFD increased. Then $FCO_2$ started to increase, becoming positive at 2000. $T_{air}$ reached a minimum of 2.2 °C at 0700 and a maximum of 19.9 °C at 1700. Thus, this alpine meadow is characterized by a large diurnal amplitude of air temperature variation. $T_{soil}$ reached a minimum
of 12.4 °C at 0800 and a maximum of 20.6 °C at 1800. SWC ranged between 0.267 and 0.286 because of daytime evapotranspiration and small precipitation on the morning on 12 August 2001. \( \Delta E \) paralleled PPFD. VPD paralleled \( T_{air} \) and reached a maximum of 15.4 hPa at 1700. \( R_{canopy} \) remained low at 100–140 s m\(^{-1}\) in the forenoon, but increased in the afternoon and reached a maximum of 168 s m\(^{-1}\) at 1800.

\( FCO_2 \) is plotted against PPFD in Fig. 3.33. In the afternoon, when soil temperature was high, \( FCO_2 \) was higher than the forenoon \( FCO_2 \).

### 3.3.4. GPP and \( R_e \)

The monthly averaged diurnal courses of \( GPP \) and \( R_e \) are shown in Fig. 3.34. \( GPP \) shows a maximum value of 15.1 \( \mu \text{mol m}^{-2} \text{s}^{-1} \) at 1200 in August (Fig. 3.34a). Large \( GPP \) occurred in July and August, when LAI and biomass were maximum, and small \( GPP \) in May, June, and September. \( R_e \) showed a maximum value of 7.0 \( \mu \text{mol m}^{-2} \text{s}^{-1} \) at 1600 in August (Fig. 3.34b), and increased in the afternoon as soil temperature increased.

The relationships between \( GPP \) and PPFD and between \( R_e \) and \( T_{soil} \) in July, when plant growth and photosynthesis were maximum, are shown in Figs. 3.35a and 3.35b. As PPFD increased, \( GPP \) increased but showed no significant difference in light response of \( GPP \) between morning and afternoon. As \( T_{soil} \) increased, \( R_e \) increased exponentially and also showed no significant difference in response to temperature between morning and afternoon values.

To assess the production efficiency of radiation use (\( RUE \)) and water use (\( WUE \)), hourly averaged \( GPPs \) are plotted against to PPFD and evapotranspiration \( E \) in Fig. 3.36. The \( GPP \) increased as irradiance PPFD increased during growing season, except for October – April. The \( GPP \) was large in the order August \( \prec \) July \( \prec \) June \( \prec \) September \( \prec \) May \( \prec \) October – April (Fig. 3.36a). The \( GPP \) increases as \( E \) increased in the same order.

### 3.4. Discussion

#### 3.4.1. Measurement accuracy of eddy covariance method
In the spectrum analyses of turbulent fluctuations in wind velocity and other scalars, the power spectra showed \(-5/3\) power slope (Fig. 3.7) and the cross-spectra showed \(-4/3\) power slope in the inertial sub-range (Fig. 3.8). The dimensionless universal functions showed \(-1/3\) slope against the atmospheric stability under strongly unstable conditions (Fig. 3.14, 15). These confirm the reasonability of turbulent flux sampling in this study. The estimation errors generated by the absence of coordinate rotation, trend removing and water vapor correction, are very small (Fig. 3.10-12) and this ensures that it is unnecessary to apply these corrections.

However, the energy imbalance, the disagreement between the energy fluxes \((H+\lambda E)\) and the available energy \((Rn-G)\), existed for whole experimental periods. The regression coefficients show the underestimation of the surface energy fluxes \((H+\lambda E)\) of 41\% (2001) and 30\% (2002) relative to active energy \((Rn-G)\) (Fig. 3.16). These slopes of regression lines are lower than the mean of 0.79 presented by Wilson et al. (2002), who compiled the \(EBR\) data of 22 sites and 50 site-years in FLUXNET, however they are within the range of 0.53 to 0.99 presented by Wilson et al. (2002). The \(EBR\) is slightly larger in summer than in winter (Fig. 3.17). This may be because total available energy, as the denominator in the \(EBR\) calculations, was much greater in the peak summer period relative to the winter period (Wilson et al., 2002). During day/night transition periods in morning and evening, since the mean value of \(Rn-G\) was close to zero, the \(EBR\) is not especially meaningful (Fig. 3.18). This pattern of a greater \(EBR\) in the afternoon relative to the morning was observed in other flux sites (Wilson et al., 2002). The \(EBR\) shows no significant dependence on the wind direction (Fig. 3.19). This shows that the \(EBR\) was not influenced by the footprint land cover. The \(EBR\) shows the convergence to near 0.7 in the daytime, when the friction velocity exceeded 0.3 m s\(^{-1}\), and the dispersion in the nighttime against to the friction velocity (Fig. 3.20). These results show the low \(EBR\) values were within a range of data from other flux measurement sites (Wilson et al., 2002), and show no obvious trend in the relationship to wind direction and friction velocity. The energy imbalance may be caused by sampling error, instrumental biases and other energy sinks as pointed out by Wilson et al. (2002).
3.4.2. Environmental controls on the diurnal change of carbon dioxide dynamics

One major process may affect the afternoon CO$_2$ exchange between the land and the atmosphere ($FCO_2$; Fig. 3.31): the increasing air and soil temperatures accelerate plant respiration and soil organic matter decomposition, and as a result suppress the net CO$_2$ uptake flux. In grassland ecosystems with low LAI, because of the low interception of sunlight by vegetation, the soil surface temperature could be very high, which will further accelerate the increase of soil CO$_2$ efflux. This is supposed by the higher values of $FCO_2$ together with higher soil temperature in the afternoon as shown in Fig. 3.33. In addition, the high VPD might cause the closure of stomata and thus suppress CO$_2$ uptake. Such a stomatal response to increased VPD has been observed in a C$_3$/C$_4$ prairie grassland (Verma et al., 1992), a black spruce forest (Jarvis et al., 1997), and a jack pine forest (Baldocchi and Vogel, 1997). In the alpine meadow, the increase of $R_{canopy}$ is correlated with the decrease of $\lambda E$ but with the increase of VPD (Fig. 3.32). The amplitude of the diurnal SWC change was small and may play a very limited role in ecosystem respiration (Fig. 3.32b). Cui et al. (2003) found out apparent photoinhibition at a high PPFD, caused by reduced electron transport rate (ETR), found out in two Asteraceae species at the Haibei alpine meadow. However, the net CO$_2$ uptake flux was not suppressed in the high PPFD over 800-1000 µmolm$^{-2}$s$^{-1}$.

Daytime $R_e$ and GPP were estimated from the regression relationship between the low nighttime respiration of plant and soil microbiota biomass and soil temperature (Fig. 3.34). Therefore, it is possible that respiration of aboveground plants is enhanced significantly by high soil temperature, or PPFD in the daytime increases $R_e$ more than expected. However, these results provide evidence that respiration by plants biomass and soil microbes greatly decreases the diurnal changes of net CO$_2$ exchange in alpine meadow ecosystems, as hypothesized above.

$R_e$ increased significantly in the afternoon (Fig. 3.34 and 3.35). In this grassland, because of the small interception of sunshine by plants, the soil surface temperature increased in the mid-afternoon was significantly higher than in the nighttime and early morning, and as a result the CO$_2$ efflux was noticeably enhanced. GPP increased nearly linearly with increasing
PPFD (Fig. 3.35), and there was no suppression of GPP flux in the afternoon, as found in net ecosystem flux of CO2 exchange.

The diurnal course of GPP response to the PPFD (Fig. 3.36a), showed smaller hysteresis than found in FCO2 (Fig. 3.31) during the growing season. Furthermore, the diurnal course of GPP response against the evapotranspiration (Fig. 3.36b), showed no hysteresis in afternoon, except for July (Fig. 3.36a). This presents that the plant photosynthesis was not suppressed by water loss through leaf stomata. Cui et al. (2002) also showed that photochemical reaction in leaves of two Asteraceae species was aggravated by high leaf temperature at the Haibei alpine meadow. However, the light response of GPP was not suppressed in the afternoon with high air and soil temperature.

3.4.3. Ecosystem carbon assimilation ability

The maximum CO2 uptake flux (minimum FCO2 value) of study site (−10.8 μmol m−2 s−1; Fig. 3.30) was compared with those of other sites at a similar latitude (Table 3.2). The maximum CO2 uptake flux was two-thirds less than those in an Oklahoma C3/C4 prairie (−15.5 μmol m−2 s−1; Sims and Bradford, 2001) and a Colorado subalpine conifer forest of subalpine fir and black spruce (−15.5 μmol m−2 s−1; 3050 m; Monson et al., 2002). But it was 20%–40% less than those in a Kansas C4 prairie (−25 μmol m−2 s−1; Ham and Knapp, 1998), an Oklahoma tall-grass prairie (−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).

The maximum CO2 release flux (maximum FCO2 value) of study site (4.4 μmol m−2 s−1; Fig. 3.30; Table 3.2) was compared similarly. It was 20%–50% less than those in the Colorado subalpine conifer forest (8–9 μmol m−2 s−1; Monson et al., 2002), the Kansas C4 prairie (10 μmol m−2 s−1; Ham and Knapp, 1998), the Oklahoma tall-grass prairie (11.4 μmol m−2 s−1; Suyker and Verma, 2001), and the Japanese C3/C4 grassland (21.6 μmol m−2 s−1; Li and Oikawa, 2001).

Thus, this Tibetan alpine meadow ecosystem shows a lower CO2 uptake and lower release potential than C4 grasslands, but a similar net CO2 uptake potential to an alpine conifer forest because of its equal uptake potential and its lower release potential.
The *K. humilis* meadow had relatively high LAI (3.1) with relatively low aboveground biomass (ca. 300 g m$^{-2}$). With a similar LAI to study site, the aboveground biomass in an Oklahoma grassland was reported to be as large as 800 g m$^{-2}$. The lower ratio of aboveground biomass to LAI in the *K. humilis* meadow seems to have been due to the low canopy height in the alpine meadow, which resulted from the relatively greater abundance of broad-leaved species and the low height of vegetation. The high LAI, on the other hand, seems to contribute to the high belowground biomass in the alpine meadow. The belowground biomass was 6-7 times higher than the aboveground biomass (e.g. 1892 g d. w. m$^{-2}$ on 11 August 2001; unpublished data). This particular allocation pattern of plant biomass may favor high soil carbon storage in the alpine ecosystem.

Although the CO$_2$ uptake was not high, the daily net ecosystem carbon gain reached a fairly high value during a sunny week in August 2001. One reason may be why the low nighttime temperature (e.g. near freezing air temperature 2.2 °C) limited ecosystem respiration. Another reason is perhaps the low maintenance respiration due to the low aboveground biomass in the alpine meadow (Table 3.2).
<table>
<thead>
<tr>
<th>Meteorological elements</th>
<th>Instruments</th>
<th>Height (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Wind velocity and air temperature</td>
<td>Sonic anemometer (CSAT/CSI)</td>
<td>2.2</td>
</tr>
<tr>
<td>CO₂, H₂O concentration</td>
<td>CO₂/H₂O infrared gas analyzer (CS-7500/CSI)</td>
<td>2.2</td>
</tr>
<tr>
<td>Short wave and long wave radiations from the sky and the ground</td>
<td>Radiometer (CNR-1/Kipp &amp; Zoen)</td>
<td>1.5</td>
</tr>
<tr>
<td>Air temperatures and humidity</td>
<td>Thermo-hygrometer (HMP45C/Vaisara)</td>
<td>2.2, 1.1</td>
</tr>
<tr>
<td>Wind speed</td>
<td>Cup anemometer (034A-L and 014A/R. M. Young)</td>
<td>2.2, 1.1</td>
</tr>
<tr>
<td>Soil moisture</td>
<td>TDR sensor (CS615/CSI)</td>
<td>-0.05, -0.2, -0.5</td>
</tr>
<tr>
<td>Soil temperature</td>
<td>Thermocouple</td>
<td>-0.025, -0.05, -0.1, -0.2, -0.3, -0.4, -0.5, -0.6, -0.7</td>
</tr>
<tr>
<td>Soil surface temperature</td>
<td>Thermistor probe (107/CSI)</td>
<td>0 (3 points)</td>
</tr>
<tr>
<td>Soil heat flux</td>
<td>Heat plate (HFT-3/CSI)</td>
<td>-0.02 (3 points)</td>
</tr>
<tr>
<td>Precipitation</td>
<td>Tipping bucket rain gage (TE525MM/CSI)</td>
<td>0.5</td>
</tr>
<tr>
<td>PPFD</td>
<td>Quantum sensor (LI-190SB/Li-Cor)</td>
<td>1.5</td>
</tr>
<tr>
<td>UV-A</td>
<td>Ultra violet radiometer (PD204A/Macam)</td>
<td>1.5</td>
</tr>
<tr>
<td>UV-B</td>
<td>Ultra violet radiometer (PD204B/Macam)</td>
<td>1.5</td>
</tr>
</tbody>
</table>
Table 3.2. Different components of carbon exchange fluxes and environmental conditions for the *Kobresia humilis* meadow and other sites at similar latitude.

<table>
<thead>
<tr>
<th>Site</th>
<th>Latitude, Logitude</th>
<th>Altitude (m a.s.l.)</th>
<th>Maximum LAI</th>
<th>Maximum aboveground biomass (g m(^{-2}))</th>
<th>Minimum FCO(_2) (µmol m(^{-2}) s(^{-1}))</th>
<th>Maximum FCO(_2) (µmol m(^{-2}) s(^{-1}))</th>
<th>Maximum NEP(_{total}) (gCm(^{-2})d(^{-1}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>C(_3) alpine meadow(^1)</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(^2)</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(^3)</td>
<td>40°02'N 105°32'W</td>
<td>3050</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(_4) prairie(^4)</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(^5)</td>
<td>39°03'N 99°32'W</td>
<td>445</td>
<td>3.2</td>
<td>1100</td>
<td>---</td>
<td>---</td>
<td>6.3</td>
</tr>
<tr>
<td>Tall-grass prairie(^6)</td>
<td>36°56'N 96°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(_3)/C(_4) grassland(^7)</td>
<td>36°06'N 140°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>

\(^1\)Qinghai-Tibetan Plateau, China (This study). \(^2\)Oklahoma, USA (Sims and Bradford, 2001). \(^3\)Colorado, USA (Monson et al., 2002). \(^4\)Kansas, USA (Ham and Knapp, 1998). \(^5\)Kansas, USA (Kim et al., 1992). \(^6\)Oklahoma, USA (Suyker and Verma, 2001). \(^7\)Tsukuba, Japan (Li and Oikawa, 2001).
Figure 3.1. A schematic diagram of the flux measurement system
Figure 3.2. A view of the experimental site (summer, 2001)
Figure 3.3. A view of the experimental site (winter, 2002)
Figure 3.4. A view of the setting of the instruments (aboveground)
Figure 3.5. A view of the setting of the instruments (belowground)
Figure 3.6. Time series of x, y, z-axis wind velocities (u, v, w), air temperature ($T_{air}$), specific humidity ($q$) and carbon dioxide concentration ($C$) at the Haibei meadow, 12:00-12:03, 13 February, 2003. Data show scalar fluctuations measured at the frequency of 20 Hz by a sonic anemometer (CSAT-3, Campbell Scientific Inc., Logan, UT, USA) and an open-path Infra-Red Gas Analyzer (CS-7500, Campbell Scientific Inc.).
Figure 3.7. Power spectrum of x, y, z-axis wind velocities (u, v, w), air temperature (T), specific humidity (q) and carbon dioxide concentration (C) at the Haibei meadow, 12:00-12:15, 13 August, 2002. Vertical axes of the figures are normalized with frequency.
Figure 3.8. Cross spectrum between z-axis wind velocities \((w)\) and air temperature \((T_a)\), specific humidity \((q)\) and carbon dioxide concentration \((C)\) at the Haibei meadow, 12:00-12:15, 13 August, 2002.
Figure 3.9. The wind direction dependent deflection of the 15 min average wind inclination from 17 July to 26 July 2002. The effect of the tower is clear in between 100 and 270 degrees of wind direction.
Figure 3.10. Comparisons of 15 min averaged fluxes of corrected sensible heat ($H_{correct}$) with uncorrected sensible heat ($H_{uncorr}$) measured from 17 July to 26 July, 2002 at the Haibei station. "Rot", "Tre" and "Vap" mean rotation, trend removing and water vapor correction, respectively.
Figure 3.11. Comparisons of 15 min averaged fluxes of corrected latent heat ($\lambda E_{\text{correct}}$) with uncorrected latent heat ($\lambda E_{\text{uncorr}}$) measured from 17 July to 26 July, 2002 at the Haibei station. "Rot", "Tre" and "Vap" mean rotation, trend removing and water vapor correction, respectively.
Figure 3.12. Comparisons of 15 min averaged fluxes of corrected CO2 ($F_{CO2\text{correct}}$) with uncorrected CO2 ($F_{CO2\text{uncorr}}$) measured from 17 July to 26 July, 2002 at the Haibeii station. “Rot”, “Tre” and “Vap” mean rotation, trend removing and water vapor correction, respectively.
Figure 3.13. Wind direction dependent deflection of the 15 min averaged $H$, $\lambda E$ and $FCO_2$ fluxes from 14 January to 31 December, 2002. Black line shows 100 point moving average.
Figure 3.14. Relationship between aerodynamic stability and dimensionless universal functions under unstable conditions from 14 January to 31 December, 2002. Data show 15 min average values.
Figure 3.15. Relationship between aerodynamic stability and dimensionless universal functions under near neutral conditions from 14 January to 31 December, 2002. Data show 15 min average values.
Figure 3.16. Daily energy balance between $H+\lambda E$ and $Rn-G$ in 2001 and 2002.
Figure 3.17. Seasonal changes of daily energy balance ratio (EBR) between $H+\lambda E$ and $Rn-G$ in 2001 and 2002.
Figure 3.18. Daily changes of energy balance ratio ($EBR$) between $H+\Delta E$ and $Rn-G$ in 2001 and 2002. Data shows monthly average values.
Figure 3.19. Wind direction dependent of the energy balance ratio ($EBR$) from 14 January to 31 December, 2002. Data shows 15 min averages. Black line shows 100-point moving averages.
Figure 3.20. Energy balance ratio ($EBR$) dependent on the friction velocity from 14 January to 31 December, 2002. Data show 15 min averages.
Figure 3.21. Diurnal changes of hourly averaged $T_{air}$ at the Haibei station in 2001 and 2002. Data show monthly averaged values.
Figure 3.22. Diurnal changes of hourly averaged $T_{soil}$ at Haibei station in 2001 and 2002. Data show monthly averaged values.
Figure 3.23. Diurnal changes of hourly averaged PPFD at the Haibei station in 2001 and 2002. Data show monthly averaged values.
Figure 3.24. Mean and standard deviation (N=5) of the aboveground biomass and the LAI (leaf area index) for the *Koresia humilis* alpine meadow in 2002.
Figure 3.25. Diurnal changes of hourly-averaged $H$ flux at the Haibei station for each month of the years 2001 and 2002, respectively.
Figure 3.26. Diurnal changes of hourly-averaged \( \lambda E \) flux at the Haibei station for each month of the years 2001 and 2002, respectively.
Figure 3.27. Diurnal courses of hourly-means of $H$, $\lambda E$ and environmental conditions from 9 to 15 August 2001. a) available radiation ($Rn-G$, ○); sensible heat flux ($H$, ◊); latent heat flux ($\lambda E$, □); b) the difference between canopy surface temperature and air temperature ($T_{surface} - T_{air}$, ×); vapor pressure deficit ($VPD$, +); c) Omega factor ($\Omega$, △); Bowen ratio ($\beta$, ●); friction velocity ($u_*$, □). The canopy surface temperatures are estimated from the upward long wave radiation data using the Stefan-Boltzmann’s law; $Rl = \varepsilon \sigma (T_{surface} + 273.15)^4$, where $Rl$ is the long wave radiation (W m$^{-2}$), $\varepsilon$ is radiation emissivity (adopted for 0.98), $\sigma$ is the Stefan-Boltzmann constant = 5.67 x 10$^8$ W m$^{-2}$ K$^{-4}$).
Figure 3.28. Relationship between the hourly-means of soil surface-air temperature difference and sensible heat flux ($H$) from 9 to 15 August 2001. Data show hourly mean values.
Figure 3.29. Relationship between hourly-means of vapor pressure deficit (VPD) and latent heat flux ($\lambda E$) from 9 to 15 August 2001.
Figure 3.30. Diurnal changes of hourly averaged $F_{CO_2}$ flux at the Haibei station in 2001 and 2002. Data show monthly averaged value.
Figure 3.31. Relationship between net carbon exchange rate ($F_{CO_2}$) and incident $PPFD$ (photosynthetic photon flux density) in each month over a growing season: August 2001 (□), September 2001 (△), May 2002 (◆), June 2002 (○), July 2002 (×), August 2002 (■), September 2002 (▲).
Figure 3.32. Diurnal courses of hourly mean CO₂ exchange flux and environmental conditions on 9-15 August 2001. a) Net carbon exchange rate (FCO₂, □); incident photosynthetic photon flux density (PPFD, ♦); b) air temperature (T_air, ●); soil temperature (T_soil, ◇); soil water content (SWC, △); c) VPD (×), Penman–Monteith canopy resistance (R_canopy, ▲), and latent heat flux (λE, +).
Figure 3.33. Relationship between hourly-means of the PPFD and CO2 flux (FCO2) from 9 to 15 August 2001.
Figure 3.34. Diurnal courses of (a) hourly mean gross primary production ($GPP$), and (b) hourly mean ecosystem respiration ($R_e$) in 2002 (May ○, June □, July ◇, August ×, September +, October–April △).
Figure 3.35. Relationship between (a) gross primary production \((GPP)\) and irradiant \(PPFD\) (photosynthetic photon flux density), and (b) ecosystem respiration \((R_e)\) and 5 cm soil temperature \((T_{soil})\) in July 2002 \((0–12\ h, \bigcirc\ \text{with solid lines,} \ 12–24\ h\ \square, \ \text{with broken lines})\). Data show hourly mean values with standard deviation \((n = 30\ \text{or} \ 31)\). The equation for predicting \(GPP\) from \(PPFD\) is \(y = 0.000292x + 0.18; \ r = 0.612\ \text{(0–12 h)}, \ 0.432\ \text{(12–24 h)};\ \text{s.e.} = 0.0102\ \text{(0–12 h)}, \ 0.00763\ \text{(12–24 h)}\). The equation for predicting \(R_e\) from \(T_{soil}\) is \(y = 0.0174\exp(0.117x); \ r = 0.594\ \text{(0–12 h)}, \ 0.726\ \text{(12–24 h)};\ \text{s.e.} = 0.000857\ \text{(0–12 h)}, \ 0.00149\ \text{(12–24 h)}\). There is no significant difference between 0-12h and 12-24h in ANOVA of \(GPP\) and \(R_e\) data \((P < 0.0001)\).
Figure 3.36. Relationship between (a) gross primary production (GPP) and incident PPFD, and (b) GPP and evapotranspiration (E) in 2002 (May ○, June □, July ◇, August ×, September +, October–April △).
Chapter 4 Seasonal changes of the CO2 exchanges

4.1. Introduction

Net ecosystem production \((NEP = -NEE)\) between the atmosphere and terrestrial ecosystems is influenced by various factors, e.g. temperature, light intensity, soil water content and phenology (Aubinet et al., 2000; Baldocchi et al., 2001; Valentini et al., 2000). Gross primary production \((GPP)\), CO2 uptake by plants via photosynthesis, and ecosystem respiration \((R_e)\), CO2 efflux by plants and soil microbes via respiration, determine the \(NEP\). Seasonal changes in \(GPP\) are affected mainly by leaf physiological activity, leaf area index \((LAI)\), weather, and growing period. \(R_e\) is composed of autotrophic respiration \((AR)\) and heterotrophic respiration \((HR)\), which are controlled by plant and soil microbe activities disparately. \(AR\) is composed of plant growth and maintenance respiration. \(HR\) is strongly regulated by soil temperature and moisture (Lloyd and Taylor, 1994; Davidson et al., 1998; Xu and Qi, 2001). Thus, the phase and amplitude of these components determine the seasonal pattern of net CO2 ecosystem exchange \((NEE)\) (Randerson et al., 1999; White et al., 1999; Cramer et al., 1999; Falge et al., 2002). By using the FLUXNET database, Falge et al. (2002) recently studied the factors that control seasonal changes in \(GPP\) and \(R_e\) in a wide range of terrestrial ecosystems. Their result showed that the seasonality of \(GPP\) was determined by the life-form (e.g. broad leaf or needle leaf), and that of \(R_e\) was affected by the climatic types that are characterized usually by temperature. Saigusa et al. (2002) reported that the maximum \(GPP\) of a cool-temperate deciduous forest, derived from a \(PPFD–GPP\) regression curve of a rectangular hyperbolic function, was associated with \(LAI\), but that light use efficiency as the initial slope of the regression curve did not show clear seasonal changes. Flanagan et al. (2002) found that \(GPP\) in northern temperate grasslands was strongly related to \(LAI\) and canopy nitrogen content. Thus, \(GPP\) and \(R_e\) seems to be affected by plant phenology and climatic seasonality in lowland ecosystems. On the other hand, these relationships in alpine ecosystems are expected to exhibit different patterns from those in lowland ecosystems, even at similar latitudes (Li and Zhou, 1998). Little evidence, however, is available for us to determine the influence of environmental
factors on the seasonal changes in carbon dynamics of alpine ecosystems. Specifically, how do $GPP$ and $R_e$ of alpine grassland change seasonally?

From August 2001 to December 2002, long-term CO$_2$, water vapor and energy flux measurements using the eddy covariance method were carried out in an alpine meadow on the Qinghai-Tibetan Plateau. Additionally, the measured net ecosystem exchanges were attempted to be divided into $GPP$ and $R_e$. The aims of this chapter are 1) to show the seasonal changes of those CO$_2$ fluxes, $NEP$, $GPP$ and $R_e$, and 2) to clarify the environmental controls on the seasonal changes of those CO$_2$ dynamics.

### 4.2. Materials and Methods

CO$_2$ and H$_2$O flux were measured by the open-path eddy covariance method from 9 August to 31 December 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 2.2 m with an open-path Infra-red gas analyzer (CS-7500, Campbell Scientific Inc.). Details are described in subsection 3.2.

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 $PPFD$ meter (LI-190SB, Li-Cor, Lincoln, NE, USA). Air temperature and humidity were measured at 2.2 m and 1.1 m above the ground with a humidity and temperature probe (HMP45C, Vaisala, Helsinki, Finland). Wind speed was measured at 2.2 m and 1.1 m above the ground with a cup anemometer (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 points with heat 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 manufactured thermocouples. Soil water content was measured at 0.05, 0.2, and 0.5 cm with time-domain reflectometry sensors (CS-615, Campbell Scientific Inc.). Soil surface temperature was measured at three points in a 1-m$^2$ 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.). Details are described in subsection 3.2.

4.2.1. NEP, GPP and $R_e$ gap-filling methods

When natural and anthropogenic impacts on the global ecosystem carbon budget are being compared among biome types, phenology, and stress conditions, the calculations usually use annual sum data of net ecosystem exchange (Valentini et al., 2000; Falge et al., 2002). However, Falge et al. (2001) found that the average data coverage during a year was only 65%, owing to system failures or data rejection. They reviewed several methods of gap-filling and applied them to data sets available from the EUROFLUX and AmeriFlux databases. They used mean diurnal variation (MDV), look-up tables (LookUp), and nonlinear regression (Regression) methods, and investigated the impact of different gap-filling methods on the annual sum of NEP. In this study, those three methods were also used to fill gaps in order to obtain annual sums of NEP. Details of the methods are reported in Falge et al. (2001).

In the MDV method, a missing datum is replaced by the mean for that time period (15 min) from adjacent days. The data windows of 7 and 14 days and two different algorithms of (a) an “independent” window and (b) a “gliding” window were chosen. In (a), for each subsequent period of data, mean diurnal variations are established to fill gaps within that period. In (b), a window of prescribed size around each gap is used to construct mean diurnal variations for gap-filling within that window.

In the LookUp method, look-up tables were created for six bimonthly periods or four seasonal periods ranging from 1 April to 30 May, 1 June to 30 September, 1 October to 30 November, and 1 December to 31 March. For the look-up table, average NEEs were compiled for 27 PPFD classes × 35 $T_{air}$ (air temperature) classes. PPFD classes consisted of 100 $\mu$mol m$^{-2}$ s$^{-1}$ intervals from 0 to 2600 $\mu$mol m$^{-2}$ s$^{-1}$ with a separate class for PPFD class = 0. Similarly, $T_{air}$ classes were defined through 2 °C intervals ranging from −35 to +34 °C.

In the Regression method, regression relationships were established between the NEP
Component, i.e. \( R_e \) and \( GPP \), and associated controlling factors (temperature and light) for every month. Missing \( R_e \) was extrapolated by using exponential regression equations (Eq. (3.48)) between measured nighttime \( R_e \) with strong turbulence (\( u_* > 0.2 \text{ m s}^{-1} \)) and soil temperature at a depth of 5 cm. Nighttime eddy covariance flux data under the condition of below the \( u_* \) threshold (Aubinet et al., 2000) (0.2 m s\(^{-1}\) in this study) were also corrected with the regression equations (Eq. (3.48)); called as "\( u_* \)-correction". \( GPP \) was extrapolated by using rectangular hyperbolic regressions of daytime \( GPP \) against \( PPFD \) with Michaelis–Menten type equation (Falge et al., 2001) for every month:

\[
GPP = \frac{\alpha GPP_{\text{SAT}} \cdot PPFD}{GPP_{\text{SAT}} + \alpha \cdot PPFD},
\]

where \( \alpha \) is the initial slope of the light–\( GPP \) curve (\( \mu\text{mol CO}_2 [\mu\text{mol photon}]^{-1} \)) and is equivalent to the quantum yield. \( GPP_{\text{SAT}} \) is \( GPP \) at light saturation (\( \mu\text{mol m}^{-2} \text{ s}^{-1} \)). Those two parameters are assigned as the monthly-specific parameters.

### 4.3. Results

#### 4.3.1. Micrometeorology

The \( PPFD \) reached the maximum of 63.9 mol m\(^{-2}\) d\(^{-1}\) in late June and then decreased gradually (Fig. 4.1a). The daily average air temperature and soil temperature ranged from 2 to 15.5 °C and 6.3 to 19.2 °C, reaching maximum of 15.5 and 19.2 °C in the middle of July, respectively (Fig. 4.1b, c). Rain fell mainly from May to September in both years (Table 4.1). During 7–14 consecutive dry days following the rainy days in 2002 (Fig. 4.1d), the soil water content (\( SWC \)) decreased gradually from 0.54 cm\(^3\) cm\(^{-3}\) (saturation) to 0.2 cm\(^3\) cm\(^{-3}\) (Fig. 4.1e), and the vapor pressure deficit (\( VPD \)) often increased to 5–6 hPa (Fig. 4.1f).

\( H \) had two peaks in April and October and \( \lambda E \) reached the maximum status in July (Fig. 4.2a, b; Table 4.2). The Bowen ratio (\( \beta \)) showed very high value in winter season and kept low value below 1.0 during the growing season (Fig. 4.2c; Table 4.2). The ratio of actual evaporation (\( \lambda E \)) measured by eddy covariance measurement to equilibrium evaporation (\( \lambda E_{eq} \)) ranged from 0.8 to 1.0 during the growing season (Fig. 4.2d). This indicates that this ecosystem
has the capacity to produce enough evaporation to meet atmospheric demand, yet remain wet. Omega factor ($\Omega$) ranged from 0.2 to 0.8 during the growing season, and the seasonal pattern was quite similar to that of $\alpha E$ (Fig. 4.2e). $R_{canopy}$ derived from Eq. (3.47) remained low at around 100 s m$^{-1}$ from 23 May (DOY143) to 17 September (DOY260) in 2002 (Fig. 4.2f). This result supposes that the soil moisture was high enough to keep the stomata of canopy plants open. After this period, canopy resistance rose gradually to 300 s m$^{-1}$. Albedo kept around 0.20 during the growing season, from May to September, and around 0.23 during other period of the year, except occasional snow (Fig. 4.2g; Table 4.2).

### 4.3.2. Carbon dioxide flux

The seasonal changes of $NEP_{daytime}$, $NEP_{ nighttime}$, and $NEP_{total}$, are the daily integrated values of CO$_2$ uptake flux ($-F_{CO2}$) as net ecosystem production for daytime, nighttime, and the whole day, respectively (Fig. 4.3). In 2002, $NEP_{daytime}$ started to increase from 23 May (DOY143), and reached its maximum status on 30 June (DOY181) and local maximum of 5.4 g C m$^{-2}$ d$^{-1}$ on 7 July (DOY188). $NEP_{daytime}$ started to decrease from 19 August (DOY231) and reached the early summer value on 30 September (DOY273). $NEP_{ nighttime}$ showed the opposite change, reaching the minimum of $-2.5$ g C m$^{-2}$ d$^{-1}$ on 19 August (DOY231). $NEP_{total}$ showed similar changes to those of $NEP_{daytime}$, and peaked at 3.9 g C m$^{-2}$ d$^{-1}$ on 7 July (DOY188). $NEP_{total}$ rarely became negative, but reached the minimum of $-1.3$ g C m$^{-2}$ d$^{-1}$ on 29 September (DOY272). The cumulative $NEP_{total}$ for the growing season from 23 May to 30 September 2002 was 153.1 g C m$^{-2}$.

The author performed a linear regression (Fig. 4.4) of $NEP_{daytime}$ (Fig. 4.3) on integrated $PPFD$. The slopes of the regression line showed similar seasonal change pattern to $LAI$ (Fig. 3.24; Table 4.3), peaking in July and then decreasing in September 2002.

To investigate the relationship between ecosystem respiration and temperature, the nighttime average $F_{CO2}$ was plotted against the nighttime average soil temperature at a depth of 5 cm (Fig. 4.5). The nighttime average $F_{CO2}$ increased exponentially as soil temperature increased. However, the nighttime average flux was scattered widely within a narrow range of
soil temperatures in each month. Next, to assess the effect of SWC on ecosystem respiration, the nighttime average $F_{CO_2}$ was plotted against SWC for five periods of low SWC after heavy rainfall in summer time (Fig. 4.6). The CO$_2$ efflux increased as SWC decreased. CO$_2$ of at least 0.94 $\mu$mol m$^{-2}$ s$^{-1}$ was released even in saturated conditions. In the 2002 periods 23 May-1 June, 11-19 August, and 24-31 August (see Fig. 4.1d, 2nd, 4th, and 5th data sequences), the nighttime average soil temperature increased from 2.8 to 4.2 °C as SWC decreased.

4.3.3. GPP and $R_e$

$GPP$ equals to the sum of $R_e$ and NEP. The estimation accuracy for GPP is thus affected by the estimation error of $R_e$. Consequently, the estimated $R_e$ was compared to the $R_e$ derived as an offset of the rectangular hyperbolic regression in the light response curves of the $F_{CO_2}$ (as shown in Suyker and Verma, 2001; eq. (1); Fig. 4.7). The regression line showed a good agreement between two $R_e$s; the slope was 1.08 and offset was 0.06 g C m$^{-2}$ d$^{-1}$.

Seasonal changes in the monthly parameters of $R_e$ and GPP regression curves; $R_{10}$, $\alpha$, and GPP$_{sat}$, in Eqs. (3.48) and (4.1), are given in Table 4.4. $R_{10}$ reached a maximum status of 2.27, 1.88 $\mu$mol m$^{-2}$ s$^{-1}$ in August and September. $R_{10}$ showed negative values in January and February, and that was assumed to be from using the constant parameter in $E_a$. However, even if daytime $R_e$ calculation has the internal errors, the very small magnitude of $R_e$ in winter don’t have so significant influence in annual $R_e$ estimation. $\alpha$ showed abrupt high value in May and reached maximum status in June to August, and kept near zero values in October to April. GPP$_{sat}$ increased from May and reached maximum status in July and August, and kept near zero values in October to April.

Seasonal changes in the 15-d moving average values of PPFD and $T_{soil}$, which strongly affect GPP and $R_e$, are given in Fig. 4.8a. PPFD shows three maximal and three minimal values during the growing season from May to September, which corresponded to three rainfall events: maximums on days of the year (DOY) 181, 207, and 232; minimum on DOY166, 190, and 217. $T_{soil}$ also shows three maximum and three minimum values during the growing season, but those dates lag by 1–13 d in comparison with those of PPFD: maximum on DOY189, 213, and 237;
minimum on DOY167, 203, and 230. \( T_{soil} \) decreased gradually after the first day of maximum (DOY189).

Seasonal changes in the 15-d moving average values of GPP to \( R_e \) are given in Fig. 4.8b. GPP started to increase on 1 May (DOY121), reaching a first maximum value of 5.90 \( \mu \)mol m\(^{-2}\) s\(^{-1}\) on 7 July (DOY188), a second maximum of 6.46 \( \mu \)mol m\(^{-2}\) s\(^{-1}\) on 2 August (DOY214) and a third maximum of 5.71 \( \mu \)mol m\(^{-2}\) s\(^{-1}\) on 19 August (DOY231). After that day, GPP decreased gradually to 1.0 \( \mu \)mol m\(^{-2}\) s\(^{-1}\) on 28 September (DOY271). \( R_e \) started to increase at the beginning of March, reaching a first maximum value of 3.48 \( \mu \)mol m\(^{-2}\) s\(^{-1}\) on 12 July (DOY193), a second maximum of 5.40 \( \mu \)mol m\(^{-2}\) s\(^{-1}\) on 8 August (DOY220), and a third maximum of 4.45 \( \mu \)mol m\(^{-2}\) s\(^{-1}\) on 24 August (DOY236). Thus, the seasonal changes of GPP show that the plant growing season ranges from May to September in this alpine meadow ecosystem, but do not correspond perfectly with those of PPFD, although the seasonal changes of \( R_e \) show a good agreement with those of \( T_{soil} \).

Figure 4.8b also shows the seasonal changes in the ratio of GPP and \( R_e \). The ratio remained mostly larger than 1 from May to September. This indicates that this ecosystem was a sink of carbon during the growing season in 2002. The maximum value of the ratio was 2.07 on 24 July (DOY205).

Figure 4.8c shows the seasonal changes in the 15-d moving average values of GPP\(_{max}\) and \( R_{emax} \), the maximum values of diurnal GPP and \( R_e \) change, which are potential values of assimilation and respiration of CO2. GPP\(_{max}\) started to increase rapidly on 1 May (DOY121), exceeded 20 \( \mu \)mol m\(^{-2}\) s\(^{-1}\) for the two months from 2 July (DOY183), and reached a maximum of 23.6 \( \mu \)mol m\(^{-2}\) s\(^{-1}\) on 31 July (DOY212). After 20 August (DOY232), when GPP was 20.4 \( \mu \)mol m\(^{-2}\) s\(^{-1}\), GPP decreased rapidly to 1.0 \( \mu \)mol m\(^{-2}\) s\(^{-1}\) on 10 October (DOY283). \( R_e \) started to increase on 1 May (DOY121), reaching a first maximum value of 8.10 \( \mu \)mol m\(^{-2}\) s\(^{-1}\) on 1 July (DOY182) and an annual maximum of 12.0 \( \mu \)mol m\(^{-2}\) s\(^{-1}\) on 25 August (DOY237), and then decreased rapidly to 1.0 \( \mu \)mol m\(^{-2}\) s\(^{-1}\) on 25 October (DOY298). Thus, the seasonal changes of GPP\(_{max}\) corresponded well with those of GPP, although the annual maximum of \( R_{emax} \) lagged two weeks behind that of \( R_e \).
The radiation use efficiency ($RUE_{GPP}$), the ratio between $GPP$ and $PPFD$ ranged from 10 to 20 mmol CO$_2$ mol photon$^{-1}$ from June to August, $RUE_{GPP}$ change with plant growth (Fig. 4.9). The water use efficiency ($WUE_{GPP}$), the ratio between $GPP$ and evapotranspiration ($E$) ranged from 2.0 to 4.0 mmol CO$_2$ mol H$_2$O$^{-1}$ from June to August, and also changes along with plant growth (Fig. 4.9).

To clarify the effect of $LAI$ on CO$_2$ assimilation potential and resource use efficiency, the relationships between $LAI$ and $GPP_{max}$, $RUE_{GPP}$ and $WUE_{GPP}$ are shown in Fig. 4.10. $GPP_{max}$, $RUE_{GPP}$, and $WUE_{GPP}$ tended to increase with the increase of $LAI$ increased.

### 4.3.4. Annual sums of net ecosystem carbon dynamics

Table 4.5 lists annual sums of $GPP$, $R_e$, and $NEP$, and daily $GPP$ and daily $NEP$ as the quotient from the division of annual $GPP$ and $NEP$ by the number of days when $NEP > 0$. Missing data are filled by the Regression method for all parameters, and by the MDV and LookUp methods for $NEP$. Gap-filling methods reduced the proportion of missing data from 33.3% to 4% (Table 4.2). The remaining missing data are almost all in the coldest part of winter, from 1 to 13 January, when there is no biological activity at all, and does not affect annual sums of fluxes. The residue of peak aboveground biomass (134.7 g C m$^{-2}$: multiplying a peak aboveground dry weight biomass 283 g d.w. m$^{-2}$ by a conversion coefficient to carbon 0.476 (unpublished data)) taken from annual $NEP$ (78.5 g C m$^{-2}$ y$^{-1}$), as annual carbon storage below ground, is $-56.2$ g C m$^{-2}$ y$^{-1}$. In the data gap-filled by the Regr. methods without $u$-correction, annual balances of $GPP$ and $R_e$ are less than those with $u$-correction.

### 4.4. Discussion

#### 4.4.1. Environmental controls on the seasonal changes of net carbon dioxide exchanges

The linear regression slopes of $NEP_{daytime}$ against $PPFD$ were higher for August 2001 than August 2002 (Fig. 4.4). There are two major processes directly involved in this difference of slope, that is, a higher CO$_2$ uptake rate and a lower ecosystem respiration could result in a
higher slope. The \textit{LAI} seems the major factor causing the increase of ecosystem CO$_2$ uptake. As the \textit{LAI} increases, a larger photosynthetic leaf area and thus a higher ecosystem CO$_2$ uptake would be expected. It is indeed true for low \textit{LAI} values (Larcher 1995). The \textit{LAI} measured on 20 August 2001 was 3.1, which was about 20\% higher than the \textit{LAI} of 2.6 obtained on 25 August 2002 (Table 4.3). A detailed assessment of the effects of soil respiration on \textit{FCO$_2$} was beyond the scope of the current work, but it should be further clarified in order to understand the underlying mechanisms for the carbon cycle in the alpine meadow.

The nighttime CO$_2$ efflux (ecosystem respiration) shows a positive exponential relationship with soil temperature (Fig. 4.5) and an apparent linear relationship with \textit{SWC} (Fig. 4.6). Similar response of nighttime CO$_2$ flux to \textit{SWC} has been observed in tundra under near-saturation (Oechel et al., 1998). At this site, an excess of \textit{SWC} beyond the optimum level (e.g., 0.3) seemed to have greatly affected CO$_2$ emission. The effect of \textit{SWC} on CO$_2$ efflux in the alpine ecosystem seems different from that reported for relatively dry grassland (Hunt et al., 2002), likely because of the different range of \textit{SWC} encountered. However, the decrease of \textit{FCO$_2$} with \textit{SWC} in Fig. 4.6 could be an artifact of changing soil temperatures whose effect was not separated in this analysis, and therefore deserves further investigation.

\textbf{4.4.2. Environmental controls on the seasonal changes of \textit{GPP} and \textit{R$_e$}}

The timings of the maximum and minimum daily mean \textit{GPP} synchronized with those of \textit{GPP$_{max}$}, an index of potential ecosystem photosynthesis (Figs. 4.8a–c). \textit{GPP$_{max}$} is positively related to \textit{LAI} (Fig. 4.10a) as shown in Saigusa et al. (2002) and Flanagan et al. (2002). \textit{RUE$_{GPP}$} and \textit{WUE$_{GPP}$} also tended to increase as \textit{LAI} increased (Figs. 4.10b, c). These results suggest that \textit{LAI} determines the ecosystem capacity for assimilation and resource requirements.

The timings of the maximum and minimum of the daily mean \textit{R$_e$} and \textit{R$_{emax}$} synchronized with those of \textit{T$_{soil}$}, an environmental factor, when these parameters were maximum (DOY180–230) (Figs. 4.8a–c). \textit{R$_{emax}$}, however, increased even though \textit{T$_{soil}$} decreased during the same period, as seen in \textit{R$_{10}$} (Table 4.4). In general, seasonal changes in respiratory processes are controlled by climate more strongly than by biological factors (Falge et al., 2002). However, at
this study site, $R_{\text{emax}}$, which reached the maximum in autumn, may be associated with above- and below-ground biomass, which reached the maximum at the same time (Kato et al., in press, 2004b; Li and Zhou, 1998).

Seasonal changes of mean GPP and $R_e$ were synchronized in the phases, as seen in temperate and Mediterranean ecosystems (Falge et al., 2002) (Fig. 4.8b). In contrast, those of $GPP_{\text{max}}$ and $R_{\text{emax}}$ were out of phase, as presented in $GPP_{\text{SAT}}$ and $R_{10}$ (Table 4.4) and seen in boreal coniferous forests (Falge et al., 2002). That time lag of $GPP_{\text{max}}$ and $R_{\text{emax}}$ phases is expected to be caused by the differences in the seasonal patterns of biological functions, i.e. LAI, plant and soil biomass, affecting the potential of assimilation and respiration, respectively.

When the GPP/Re ratio was assumed to be conservative in an alpine meadow as suggested in forests (0.47 ± 0.04 s.d.; Waring et al., 1998), the relatively high ratio of GPP/Re indicates a high contribution of photosynthesis and autotrophic respiration to the carbon dynamics of this ecosystem (Fig. 4.8b). Values of GPP/Re near 2 in late July correspond to NEP = $R_e$, indicating a low overall contribution of HR. This suggests that mainly autotrophic processes govern ecosystem carbon fluxes, as was observed in field crops and at temperate sites (Falge et al., 2002). When GPP considerably exceeds $R_e$, like this, the accumulation of litter can reduce free nutrients in the ecosystem. Considering the close link between soil organic matter decomposition and nutrient cycling, this system may show negative feedback in growth and CO2 assimilation or be susceptible to disturbance (Shulze et al., 1999; Amiro, 2001).

### 4.4.3. Ecosystem carbon assimilation ability

The author compared the maximum NEP$_{\text{total}}$ of study site (3.9 g C m$^{-2}$ d$^{-1}$) with those of other sites at similar latitudes (Table 3.2). The maximum NEP$_{\text{total}}$ in the present study was slightly less than that of the Kansas C$_4$ prairie (6.3 g C m$^{-2}$ d$^{-1}$, Kim et al., 1992; 4.9 g C m$^{-2}$ d$^{-1}$, Ham and Knapp, 1998), but 50% less than that of the Oklahoma tall-grass prairie (8.4 g C m$^{-2}$ d$^{-1}$, Suyker and Verma, 2001). It seems that the alpine meadow ecosystem has a smaller daily CO2 uptake potential than other grassland ecosystems. The daily uptake potential of this site was 4 times that in the Colorado subalpine conifer forest at approximately the same elevation.
The low temperature may suppress plant respiration and decomposition of soil organic matter. The small biomass may also an important factor decreasing plant respiration significantly. On the other hand, the short growing season was compensated by the high daily CO$_2$ uptake potential. Consequently, the alpine ecosystem sequestered a large amount of CO$_2$ (153 g C m$^{-2}$) during the growing season in 2002. Although it is still under debate whether or not CO$_2$ fluxes should be corrected by the magnitude of energy budget closure, the simple correction that compensate the lack proportion of energy balance as presented by Saigusa et al. (2002), was applied: 226 g C m$^{-2}$ in the growing season CO$_2$ budget.

The author compared the GPP of the alpine meadow with those of other ecosystems. The annual GPP of 575 g C m$^{-2}$ was lower than those of boreal coniferous forests (723–959 g C m$^{-2}$ y$^{-1}$) and Colorado subalpine coniferous forest (831 g C m$^{-2}$ y$^{-1}$; 3050 m) at similar elevations, and much lower than that of tropical forest (3249 g C m$^{-2}$ y$^{-1}$), but within the range for temperate ecosystems, including forests and grasslands (542–1924 g C m$^{-2}$ y$^{-1}$; average, 1262 g C m$^{-2}$ y$^{-1}$; Falge et al., 2002). The daily GPP of the study site (3.59 g C m$^{-2}$ d$^{-1}$, Table 4.5) was similar to those of boreal evergreen forest and Colorado subalpine coniferous forest (4.6 and 4.4 g C m$^{-2}$ d$^{-1}$, respectively), although slightly lower than those of temperate coniferous forest and C$_3$ crops and grassland (5.7–6.9 g C m$^{-2}$ d$^{-1}$). Thus, although the alpine meadow ecosystem has a daily CO$_2$ assimilation equal to that of a Colorado subalpine forest ecosystem, it has a lower annual GPP because of the restriction of the growing period.

The annual NEP in our study site (78.5 g C m$^{-2}$ y$^{-1}$, Table 4.5), which was gap-filled by the Regr. method using the $u_-$-correction reported by Falge et al. (2002), was close to that of the Colorado subalpine coniferous forest (71 g C m$^{-2}$ y$^{-1}$), although substantially lower than that of lowland grassland (231.3 g C m$^{-2}$ y$^{-1}$) and boreal ecosystems (121.4 g C m$^{-2}$ y$^{-1}$). The daily NEP of this site (0.49 g C m$^{-2}$ d$^{-1}$, Table 4.5) was similar to that of the Colorado subalpine forest (0.38 g C m$^{-2}$ d$^{-1}$). Although this alpine meadow ecosystem has a lower annual GPP than that of the subalpine forest ecosystem, it has a comparable annual NEP. It is assumed that not
only low temperature but also small biomass suppresses the ecosystem respiration; as a result, this ecosystem may sequester substantial C.

4.4.4. Gap-filling methods and annual carbon dynamics

Annual NEPs differed largely with some gap-filling methods (78.5–149.5 g C m⁻²; Table 4.5). For the Regression method, which calculates GPP and Rₑ separately, Rₑ with u--correction was higher by 83.1 g C m⁻² y⁻¹ than without u--correction. This fact, based on the principle that u--correction replaces nighttime eddy covariance fluxes under more stable condition with fluxes under higher turbulence, should be reasonable. The consequent decline in annual NEP due to u--correction (−35.4 g C m⁻² y⁻¹) was lower than that observed in other ecosystems (−77.0 g C m⁻² y⁻¹; Falge et al., 2001). Among the three gap-filling methods, MDV gave the highest annual NEP. This fact suggests that environmental factors during gap-periods have potential to be biased by the extreme climate. The differences among the gap-filling methods were smaller than those among various ecosystem types (Falge et al., 2001). It is not able to answer conclusively which method is best. However, when the annual NEP estimated from gap-filled data is used to evaluate an ecosystem’s capacity for CO₂ assimilation or for comparison with other ecosystems, the reduction of errors should be considered.
Table 4.1. Monthly meteorological conditions from August 2001 to December 2002 in Haibei, Qinghai, China.

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<td>°C</td>
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<td>5.2</td>
<td>4.7</td>
<td>---</td>
<td>0.30</td>
<td>0.32</td>
<td>0.67</td>
<td>0.22</td>
<td>17.5</td>
<td>0.25</td>
</tr>
<tr>
<td>2002</td>
<td>Sum.</td>
<td>12037.1</td>
<td>---</td>
<td>---</td>
<td>---</td>
<td>450.0$^d$</td>
<td>---</td>
<td>---</td>
<td>---</td>
<td>---</td>
<td>6405.6$^e$</td>
<td>---</td>
</tr>
</tbody>
</table>

$T_{air}$, $T_{soil}$ 5 cm, $T_{soil}$ 50 cm, SWC 5 cm, SWC 50 cm, relative humidity (RH), VPD and albedo show monthly averaged values. PPFD and solar radiation (Rs) show daily summed values. Precipitation (Prec) shows monthly summed value. $^*$ TDR probes were assumed to indicate no meaningful values for frozen soil moisture.

$^b$ Weighted averages are conducted for days in each month. $^c$ A unit is in mol photon m$^{-2}$ yr$^{-1}$. $^d$ a unit is in mm yr$^{-1}$. $^e$ a unit is in MJ m$^{-2}$ yr$^{-1}$. 
Table 4.2. Monthly heat budgets and carbon dynamics from August 2001 to December 2002 in Haibei, Qinghai, China.

<table>
<thead>
<tr>
<th>Year</th>
<th>Month</th>
<th>$H$</th>
<th>$\lambda E$</th>
<th>$\beta$</th>
<th>$Rn$</th>
<th>$G$</th>
<th>$EBR$</th>
<th>$FCO_2^a$</th>
<th>$NEP^b$</th>
<th>$GPP^b$</th>
<th>$R_e^b$</th>
<th>Gap(un-corrected)</th>
<th>Gap (corrected)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>MJ m$^2$ day$^{-1}$</td>
<td>MJ m$^2$ day$^{-1}$</td>
<td>Dimensi-onless</td>
<td>MJ m$^2$ day$^{-1}$</td>
<td>Dimensi-onless</td>
<td>g C m$^2$ mon$^{-1}$</td>
<td>g C m$^2$ mon$^{-1}$</td>
<td>g C m$^2$ mon$^{-1}$</td>
<td>g C m$^2$ mon$^{-1}$</td>
<td>%</td>
<td>%</td>
<td></td>
</tr>
<tr>
<td>2001</td>
<td>Aug.</td>
<td>2.21</td>
<td>5.19</td>
<td>0.43</td>
<td>11.52</td>
<td>0.08</td>
<td>0.65</td>
<td>-56.9</td>
<td>---</td>
<td>---</td>
<td>---</td>
<td>---</td>
<td>---</td>
</tr>
<tr>
<td></td>
<td>Sep.</td>
<td>2.28</td>
<td>3.24</td>
<td>0.70</td>
<td>8.83</td>
<td>-0.10</td>
<td>0.62</td>
<td>-19.9</td>
<td>---</td>
<td>---</td>
<td>---</td>
<td>---</td>
<td>---</td>
</tr>
<tr>
<td></td>
<td>Oct.</td>
<td>2.59</td>
<td>0.92</td>
<td>2.80</td>
<td>5.49</td>
<td>-0.29</td>
<td>0.61</td>
<td>15.9</td>
<td>---</td>
<td>---</td>
<td>---</td>
<td>---</td>
<td>---</td>
</tr>
<tr>
<td></td>
<td>Nov.</td>
<td>2.17</td>
<td>0.39</td>
<td>5.62</td>
<td>2.74</td>
<td>-1.03</td>
<td>0.68</td>
<td>8.7</td>
<td>---</td>
<td>---</td>
<td>---</td>
<td>---</td>
<td>---</td>
</tr>
<tr>
<td></td>
<td>Dec.</td>
<td>1.72</td>
<td>0.23</td>
<td>7.62</td>
<td>1.78</td>
<td>-1.74</td>
<td>0.55</td>
<td>-4.6</td>
<td>---</td>
<td>---</td>
<td>---</td>
<td>---</td>
<td>---</td>
</tr>
<tr>
<td>2002</td>
<td>Jan.</td>
<td>1.32</td>
<td>0.38</td>
<td>3.45</td>
<td>2.11</td>
<td>-1.46</td>
<td>0.48</td>
<td>-1.2</td>
<td>1.5</td>
<td>-0.4</td>
<td>-2.0</td>
<td>50.4</td>
<td>44.2</td>
</tr>
<tr>
<td></td>
<td>Feb.</td>
<td>2.63</td>
<td>0.36</td>
<td>7.35</td>
<td>4.25</td>
<td>-0.37</td>
<td>0.64</td>
<td>1.9</td>
<td>-2.7</td>
<td>-3.1</td>
<td>-0.4</td>
<td>4.0</td>
<td>0.8</td>
</tr>
<tr>
<td></td>
<td>Mar.</td>
<td>3.28</td>
<td>0.64</td>
<td>5.09</td>
<td>6.64</td>
<td>0.40</td>
<td>0.63</td>
<td>8.3</td>
<td>-9.7</td>
<td>-4.8</td>
<td>4.9</td>
<td>13.0</td>
<td>0.0</td>
</tr>
<tr>
<td></td>
<td>Apr.</td>
<td>3.45</td>
<td>1.11</td>
<td>3.10</td>
<td>8.28</td>
<td>0.99</td>
<td>0.63</td>
<td>17.0</td>
<td>-22.7</td>
<td>-6.7</td>
<td>16.0</td>
<td>30.8</td>
<td>0.5</td>
</tr>
<tr>
<td></td>
<td>May</td>
<td>2.89</td>
<td>2.88</td>
<td>1.00</td>
<td>10.48</td>
<td>0.78</td>
<td>0.60</td>
<td>4.1</td>
<td>-9.0</td>
<td>26.6</td>
<td>35.7</td>
<td>44.4</td>
<td>0.0</td>
</tr>
<tr>
<td></td>
<td>Jun.</td>
<td>2.41</td>
<td>5.43</td>
<td>0.44</td>
<td>12.85</td>
<td>1.05</td>
<td>0.66</td>
<td>-36.6</td>
<td>34.4</td>
<td>119.9</td>
<td>85.5</td>
<td>27.1</td>
<td>0.0</td>
</tr>
<tr>
<td></td>
<td>Jul.</td>
<td>2.25</td>
<td>7.03</td>
<td>0.32</td>
<td>12.88</td>
<td>0.59</td>
<td>0.76</td>
<td>-84.6</td>
<td>84.3</td>
<td>184.4</td>
<td>100.1</td>
<td>22.3</td>
<td>0.0</td>
</tr>
<tr>
<td></td>
<td>Aug.</td>
<td>2.30</td>
<td>6.40</td>
<td>0.36</td>
<td>12.04</td>
<td>0.13</td>
<td>0.73</td>
<td>-54.8</td>
<td>30.3</td>
<td>188.3</td>
<td>158.0</td>
<td>29.6</td>
<td>0.0</td>
</tr>
<tr>
<td></td>
<td>Sep.</td>
<td>2.67</td>
<td>3.12</td>
<td>0.86</td>
<td>8.75</td>
<td>-0.09</td>
<td>0.65</td>
<td>-8.9</td>
<td>-8.9</td>
<td>-1.6</td>
<td>73.0</td>
<td>74.6</td>
<td>12.8</td>
</tr>
<tr>
<td></td>
<td>Oct.</td>
<td>3.56</td>
<td>1.36</td>
<td>2.62</td>
<td>6.66</td>
<td>-0.24</td>
<td>0.71</td>
<td>2.0</td>
<td>-15.2</td>
<td>5.2</td>
<td>20.4</td>
<td>78.6</td>
<td>0.0</td>
</tr>
<tr>
<td></td>
<td>Nov.</td>
<td>1.97</td>
<td>0.44</td>
<td>4.50</td>
<td>2.19</td>
<td>-0.97</td>
<td>0.76</td>
<td>4.5</td>
<td>-6.8</td>
<td>-5.2</td>
<td>1.6</td>
<td>22.2</td>
<td>0.5</td>
</tr>
<tr>
<td></td>
<td>Dec.</td>
<td>1.77</td>
<td>0.26</td>
<td>6.76</td>
<td>1.60</td>
<td>-1.14</td>
<td>0.74</td>
<td>1.3</td>
<td>-4.3</td>
<td>-2.1</td>
<td>2.3</td>
<td>13.5</td>
<td>0.0</td>
</tr>
<tr>
<td>2002</td>
<td>Ave.$^c$</td>
<td>2.54</td>
<td>2.43</td>
<td>1.05$^d$</td>
<td>7.37</td>
<td>-0.03</td>
<td>0.67</td>
<td>-12.2</td>
<td>6.1</td>
<td>47.2</td>
<td>41.1</td>
<td>29.2</td>
<td>3.9</td>
</tr>
<tr>
<td>2002</td>
<td>Sum.</td>
<td>927.96$^e$</td>
<td>885.79$^e$</td>
<td>---</td>
<td>2689.09$^e$</td>
<td>-11.88$^e$</td>
<td>---</td>
<td>-146.4$^f$</td>
<td>78.5$^f$</td>
<td>575.1$^f$</td>
<td>496.6$^f$</td>
<td>---</td>
<td>---</td>
</tr>
</tbody>
</table>

*H, $\lambda E, Rn$ and $G$ show monthly averaged values. $FCO_2$, $NEP$, $GPP$ and $R_e$ show daily summed values. $^a$ This is not a gap-filled value. $^b$ These are gap-filled values. $^c$ Weighted averages are conducted for days in each month. $^d$ This is the ratio between annual summed $H$ and $\lambda E$. * units are in MJ m$^2$ yr$^{-1}$. $^f$ units are in g C m$^2$ yr$^{-1}$. 

87
Table 4.3. Aboveground biomass and leaf area index (LAI) in 2002 in Haibei, Qinghai, China.

<table>
<thead>
<tr>
<th>Date</th>
<th>Aboveground Biomass</th>
<th>LAI</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>g D.W. m(^2)</td>
<td>m(^2) m(^2)</td>
</tr>
<tr>
<td>25-May</td>
<td>49.7</td>
<td>---</td>
</tr>
<tr>
<td>31-May</td>
<td>48.5</td>
<td>---</td>
</tr>
<tr>
<td>13-Jun</td>
<td>79.2</td>
<td>0.53</td>
</tr>
<tr>
<td>30-Jun</td>
<td>107.5</td>
<td>1.1</td>
</tr>
<tr>
<td>16-Jul</td>
<td>205.0</td>
<td>3.8</td>
</tr>
<tr>
<td>30-Jul</td>
<td>283.4</td>
<td>2.7</td>
</tr>
<tr>
<td>13-Aug</td>
<td>260.6</td>
<td>3.1</td>
</tr>
<tr>
<td>25-Aug</td>
<td>280.0</td>
<td>2.6</td>
</tr>
<tr>
<td>10-Sep</td>
<td>194.5</td>
<td>2.3</td>
</tr>
</tbody>
</table>

Table 4.4. Parameterization of ecosystem respiration (\(R_e\)) and gross primary production (GPP) regression curves using 15 min averaged data in 2002.

<table>
<thead>
<tr>
<th>Month</th>
<th>(\frac{R_e}{R_{10}}) µmol CO(_2) m(^2) s(^{-1})</th>
<th>(r^2)</th>
<th>(\alpha) (\times 10^3) µmol CO(_2) µmol photon(^{-1})</th>
<th>GPP (\mu)mol CO(_2) m(^2) s(^{-1})</th>
<th>(GPP_{Sat}) (\mu)mol CO(_2) m(^2) s(^{-1})</th>
<th>(r^2)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Jan.</td>
<td>-0.75</td>
<td>0.003</td>
<td>0.86</td>
<td>-0.02</td>
<td>0.004</td>
<td></td>
</tr>
<tr>
<td>Feb.</td>
<td>-0.03</td>
<td>0.004</td>
<td>0.12</td>
<td>-0.01</td>
<td>0.089</td>
<td></td>
</tr>
<tr>
<td>Mar.</td>
<td>0.52</td>
<td>0.007</td>
<td>0.36</td>
<td>-0.02</td>
<td>0.076</td>
<td></td>
</tr>
<tr>
<td>Apr.</td>
<td>0.93</td>
<td>0.008</td>
<td>0.12</td>
<td>-0.01</td>
<td>0.120</td>
<td></td>
</tr>
<tr>
<td>May</td>
<td>1.31</td>
<td>0.018</td>
<td>139.50</td>
<td>1.54</td>
<td>0.004</td>
<td></td>
</tr>
<tr>
<td>Jun.</td>
<td>1.50</td>
<td>0.026</td>
<td>43.99</td>
<td>10.46</td>
<td>0.137</td>
<td></td>
</tr>
<tr>
<td>Jul.</td>
<td>1.23</td>
<td>0.006</td>
<td>31.83</td>
<td>20.37</td>
<td>0.489</td>
<td></td>
</tr>
<tr>
<td>Aug.</td>
<td>2.27</td>
<td>0.009</td>
<td>48.56</td>
<td>18.29</td>
<td>0.446</td>
<td></td>
</tr>
<tr>
<td>Sep.</td>
<td>1.88</td>
<td>0.124</td>
<td>24.47</td>
<td>8.20</td>
<td>0.276</td>
<td></td>
</tr>
<tr>
<td>Oct.</td>
<td>1.26</td>
<td>0.440</td>
<td>-2.32</td>
<td>0.10</td>
<td>0.119</td>
<td></td>
</tr>
<tr>
<td>Nov.</td>
<td>0.21</td>
<td>0.000</td>
<td>-1.36</td>
<td>-0.88</td>
<td>0.055</td>
<td></td>
</tr>
<tr>
<td>Dec.</td>
<td>0.44</td>
<td>0.009</td>
<td>-2.33</td>
<td>-0.22</td>
<td>0.007</td>
<td></td>
</tr>
</tbody>
</table>

The regression follows an Arrhenius-type exponential relationship for \(R_e\) (nighttime \(FCO_2\) (\(u, > 0.2\) m s\(^{-1}\))): \(R_e = R_{10} \exp\left(E_a/R \left(1/283.15 - 1/(T_{soil} + 273.15)\right)\right)\), where \(R_{10}\) is the respiration rate at the soil temperature of 10 °C (µmol CO\(_2\) m\(^2\) s\(^{-1}\)), \(E_a\) is the active energy (is adopted for only one value for whole year; 81519 J mol\(^{-1}\)), \(R\) is the gas constant (= 8.134 J K\(^{-1}\) mol\(^{-1}\)), and a rectangular hyperbolic relationship for GPP: \(GPP = \alpha \times GPP_{Sat} \times PPFD / (GPP_{Sat} + \alpha \times PPFD)\), where \(\alpha\) is quantum yield (dimensionless), and \(GPP_{Sat}\) is the saturated GPP (µmol CO\(_2\) m\(^2\) s\(^{-1}\)).
Table 4.5. Ecosystem carbon dynamics in 2002 in Haibei, Qinghai, China.

<table>
<thead>
<tr>
<th>Method</th>
<th>Specification</th>
<th>Average period</th>
<th>GPP (gCm⁻²)</th>
<th>Re (gCm⁻²)</th>
<th>NEP (gCm⁻²)</th>
<th>Days (NEP&gt;0) (days)</th>
<th>daily GPP (gCm⁻²)</th>
<th>daily NEP (gCm⁻²)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Regr.</td>
<td>with u-correction</td>
<td>1 month</td>
<td>575.1</td>
<td>496.6</td>
<td>78.5</td>
<td>160</td>
<td>3.59</td>
<td>0.49</td>
</tr>
<tr>
<td></td>
<td>without u-correction</td>
<td>1 month</td>
<td>527.4</td>
<td>413.5</td>
<td>113.9</td>
<td>164</td>
<td>3.22</td>
<td>0.69</td>
</tr>
<tr>
<td>MDV</td>
<td>independent-window</td>
<td>7 days</td>
<td></td>
<td></td>
<td></td>
<td>147.2</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>14 days</td>
<td></td>
<td></td>
<td></td>
<td>145.2</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>gliding-window</td>
<td>7 days</td>
<td></td>
<td></td>
<td></td>
<td>149.5</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>14 days</td>
<td></td>
<td></td>
<td></td>
<td>141.8</td>
<td></td>
<td></td>
</tr>
<tr>
<td>LookUp</td>
<td>T&lt;sub&gt;soil&lt;/sub&gt;</td>
<td>2 months</td>
<td></td>
<td></td>
<td></td>
<td>122.3</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Seasonal</td>
<td></td>
<td></td>
<td></td>
<td>96.9</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>T&lt;sub&gt;air&lt;/sub&gt;</td>
<td>2 months</td>
<td></td>
<td></td>
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<td>142.3</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Seasonal</td>
<td></td>
<td></td>
<td></td>
<td>113.8</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

"Day (NEP > 0)" means the number of days when daily NEP > 0 (length of growing season). GPP, Re, and NEP mean the periodic sums of gross primary production, ecosystem respiration, and net ecosystem production, respectively. Daily GPP and daily NEP were calculated by dividing GPP, Re, and NEP by the length of the growing season.
Figure 4.1. Seasonal changes in meteorological conditions in 2001 and 2002 in Haibei, Qinghai, China in 2001 and 2002. a) Photosynthetic photon flux density (PPFD); b) air temperature; c) soil temperature at 5-cm depth; d) precipitation; e) soil water content at 5-cm depth (SWC) (1, 18–28 Aug 2001; 2, 23 May–1 Jun 2002; 3, 7–18 Jun 2002; 4, 11–19 Aug 2002; 5, 24–31 Aug 2002); f) water vapor pressure deficit (VPD). Gray zones (① ~ ⑤) correspond to soil water content (SWC) shown in Fig. 4.6.
Figure 4.2. Seasonal changes in daily heat fluxes, moisture and canopy physiological conditions in Haibei, Qinghai, China in 2001 and 2002. a) sensible heat flux; b) latent heat flux; c) the Bowen ratio (\(\beta\)); d) the ratio between latent heat flux (\(\lambda E\)) and equilibrium evaporation (\(\lambda E_{eq}\)); e) omega factor (\(\Omega\)); f) Penman–Monteith canopy resistance (\(R_{canopy}\)) and f) albedo.
Figure 4.3. Changes in daily net ecosystem production in 2001 (upper graph) and 2002 (lower graph). $NEP_{daytime}$, $NEP_{nighttime}$, and $NEP_{total}$ are the sums of the net CO$_2$ uptake flux ($-F_{CO_2}$) for daytime, nighttime, and all day, respectively.
Figure 4.4. Linear regression of daytime accumulated CO₂ uptake flux ($NEP_{daytime}$) on incident photosynthetic photon flux density ($PPFD$). The regression follows a linear relationship: $NEP_{daytime} = a \times PPFD$. Monthly values are presented as follows: Month (symbol, $a$, $r^2$). August 2001 (□; 0.0793, 0.418 **), September 2001 (△; 0.0556, 0.324 **), May 2002 (●; 0.00671, 0.0653 ns), June 2002 (●; 0.0398, 0.170 *), July 2002 (×; 0.0773, 0.0441 ns), August 2002 (■; 0.0641, 0.222 *), and September 2002 (▲; 0.0322, 0.0344 ns). The linear relationships were significant at $p<0.05$ *, $p<0.01$ ** levels of correlation coefficients. ns shows nonsignificant linear relationships.
Figure 4.5. Relationship between nighttime CO$_2$ flux density and nighttime soil temperature at the depth of 5 cm in 2001 and 2002. Data show nighttime averaged values. The exponential relation is described by $FCO_2 = 0.513 \exp(0.117 \times T_{soil})$, $r^2 = 0.75$, $p<0.0001$. 
Figure 4.6. Relationship between nighttime CO$_2$ flux density and nighttime soil water content (SWC) during a dry spell after heavy rainfall (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 $F_{CO_2} = -9.97 \times SWC + 6.77$, $r^2 = 0.345$, $p<0.0001$. 

97
Figure 4.7. Ecosystem respiration ($R_e$), calculated from light response relationships (Eq. (1) in Suyker and Verma (2001)) compared to the values derived from exponential regressions between soil temperature and nighttime fluxes under turbulent conditions (Eq. (3.48)). Data represent average daily sums of each month on which $R_e$ and GPP regression curves are made in 2002.
Figure 4.8. Seasonal changes in (a) daily mean photosynthetic photon flux density (PPFD; solid lines) and 5 cm soil temperature ($T_{soil}$; broken lines); (b) daily mean gross primary production (GPP; solid lines), ecosystem respiration ($R_e$; broken lines), and ratio between GPP and $R_e$ (dashed lines); and (c) daily maximum GPP (solid lines) and $R_e$ (broken lines) in 2002. Data represent the daily means of 15-d moving average GPP/$R_e$ ratio from 14 January to 31 December, 2002.
Figure 4.9. Changes in radiation- and water-use efficiency of gross primary production in $RUE_{GPP}$ (upper graph) and $WUE_{GPP}$ (lower graph) in 2002. $RUE_{GPP}$, $WUE_{GPP}$ are the ratio of the $GPP$ to the incident $PPFD$ and evapotranspiration, respectively.
Figure 4.10. LAI controls on (a) maximum gross primary production ($GPP_{max}$), (b) radiation use efficiency ($RUE_{GPP}$), and (c) water use efficiency ($WUE_{GPP}$) of $GPP$. Vertical axis data show daily mean value of less than 4 d backward and forward LAI sampling day. $GPP_{max}$ data when the PPFD was below 1500 μmol m$^{-2}$ s$^{-1}$ were eliminated from the average. Error bars mean standard deviation (horizontal axis: $n = 5$, vertical axis: $n = 7$). In the equation for predicting $GPP_{max}$, $RUE_{GPP}$, and $WUE_{GPP}$ from LAI, standard errors are 1.560, 0.913, 0.177, respectively; $P < 0.0001$, 0.0001, and 0.05.

5.1. Introduction

Sim-CYCLE is a mechanistic model, on the basis of the dry-matter production theory established by Monsi and Saeki (1953). The atmosphere-biosphere CO2 exchange is composed of physiological processes, such as photosynthesis (GPP), respiration (AR), and decomposition (HR), and then this simulator enables us to estimate the ecosystem C budget in a mechanistic way. Plant growth process was retrieved by the Monsi (1960) scheme, and C dynamics including soil organic matter was captured by the Oikawa’s (1985) compartment model. Then, terrestrial ecosystems were conceptualized as a five-compartment system: foliage, stem and branch, root, litter, and mineral soil. In addition to GPP, AR, and HR, carbon flows such as litterfall, photosynthate allocation, and humus formation were properly formulated. GPP was regulated by a couple of physiological processes at a single-leaf scale (e.g. stomatal conductance and quantum yield), and scaled up to a canopy scale. AR consists of two functional components, i.e. maintenance respiration and growth respiration, each of which is regulated by environmental factors independently. The difference between GPP and AR is termed net primary production (NPP), and the difference between NPP and HR is termed net ecosystem production (NEP). Sim-CYCLE contains water and radiation subschemes to estimate physical environment in terrestrial ecosystems.

Ito and Oikawa (2000) performed a model analysis of the effect of climatic perturbations from 1970 to 1997 on the carbon budget of terrestrial ecosystems at the global scale, using Sim-CYCLE. During the 28 yr experimental period, global NEP ranged from -2.06 Pg C yr⁻¹ (source) in 1983 to +2.25 (sink) Pg C yr⁻¹ in 1971, being sufficiently large to give rise to anomalies in the atmospheric CO2 concentration from +0.97 to -1.06 ppmv. Regression analyses demonstrated the following: (1) annual ΔNEPs had the highest correlation ($r^2 = 0.38$) with the temperature anomaly at the global scale; (2) the anomalies in precipitation resulted in a considerable ΔNEP in northern high and middle regions; (3) an anomalous global warming by
+1°C brought about a negative $\Delta NEP$ of -2.7 Pg C yr$^{-1}$; (4) the responsiveness was primarily attributable to the temperature sensitivities of plant respiration and soil decomposition, and secondarily to the moisture sensitivity of decomposition; and (5) the temperature dependence of $\Delta NEP$ had a clear seasonality, i.e. most sensitive from July to September (summer in the northern hemisphere) relative to other seasons. In 1983, when an ENSO event happened and the tropical zone was anomalously hot (0.4 °C above the long-term mean), the largest negative $\Delta NEP$ (-2.06 Pg C yr$^{-1}$) was estimated. On the other hand, in 1971 when global mean temperature was relatively low (0.2°C below the long-term mean), the largest positive $\Delta NEP$ (+2.25 Pg C yr$^{-1}$) was estimated. Furthermore, in 1992 when an anomalous cooling during the growing period (0.3°C below the long-term mean) was caused by the Mt. Pinatubo eruption (June 1991), a considerable positive $\Delta NEP$ (+1.14 Pg C yr$^{-1}$) was estimated. The climate dependencies of global terrestrial ecosystems analyzed here may contain significant implications not only for the present functioning of atmosphere-biosphere carbon exchange, but also for ongoing global warming.

Frozen soil has a large apparent heat capacity due to the latent energy of freezing and thawing, and act as an impermeable layer because of the large hydraulic conductivity when frozen. Frozen ground, covering mid- to high-latitude regions widely, affects not only the regional scale energy and hydrological condition in those area, but also the continental scale energy and water circulation in the world. A General circulation model (GCM) simulation showed that soil freezing elevates the ground surface temperature in the mid- and high-latitude region in summer (Takata and Kimoto, 2000). The higher temperature over land results from lower evaporation caused by lower surface soil moisture, and the lower surface soil moisture is caused by additional runoff of snowmelt in spring due to the impermeability of frozen soil and low soil liquid water due to the lower evaporation. Thus, soil freezing gives the thermal and hydrological impacts on the climate system, and consequently the carbon dynamics on the ecosystem. In this study, the estimation schemes of soil freezing were adjoined into basic schemes in the Sim-CYCLE newly.
Water budget, e.g. evaporation, transpiration, runoff, infiltration and precipitation, determines the soil water content and plays an important role on the water availability for plants. When ground surfaces are sparsely covered with vegetation, estimation of evaporation and transpiration using the energy transfer model is difficult, because that evaporation from the soil surface beneath the crop canopy cannot be negligible in quantity, and it is necessary to treat the vapor flux from the crop fields as the total of soil and plant fluxes, transpiration from leaf stomata, and evaporation from the soil surface. However, the precedent Sim-CYCLE did not take sparse vegetation cover into account. Shuttleworth and Wallace (1985) designed a compartment model (S-W model) to estimate the evaporation from soil and transpiration from crops separately. This model accuracy also was examined by Kato et al. (in press, 2004a) in the sorghum fields, Tottori, Japan, and its availability for sparsely vegetated ground surface was confirmed. This study adopted this compartment model and adjoined into the basic schemes in the Sim-CYCLE as new subschemes. In precedent Sim-CYCLE, the runoff is parameterized by using the Bucket model (Manabe, 1969). This model has the simple calculation schemes that the storage water over a bucket capacity outflows as the runoff, and used widely in the GCM simulation. On the other hand, the Tank model (Sugawara and Maruyama, 1952; Sugawara, 1961), used widely in the flood forecast, has the complex calculation schemes and many parameters to control the runoff from multi-poles installed at the tank sidewall, however, also has a high estimation accuracy. Kondo (1993) developed “a new bucket model”, gathering both the simplicity of the Bucket model and the higher accuracy of the Tank model, to calculate the infiltration into the deeper soil layer in the soil surface having the heterogeneity in its soil properties and thickness. This model expresses the runoff in the hyperbolic approximate equation parameterized by precipitation, and applicable for various time scale estimation, from one hour to one month. This study adjoined this new bucket model into the basic schemes in the precedent Sim-CYCLE.

For twenty years, 1981-2000, the monthly carbon fluxes and carbon storages were simulated using the Sim-CYCLE at the Haibei station on the Qinghai-Tibetan Plateau, China. The aims of this chapter are 1) to show the interannual patterns of CO2 flux, 2) to clarify the
climate controls on the ecosystem carbon dynamics in the alpine meadow ecosystem, and 3) to predict the ecosystem response of the carbon sequestration against the global warming.

5.2. Model description

5.2.1. Sim-CYCLE -Basic model-

5.2.1.1. Overview of Sim-CYCLE

Sim-CYCLE was developed based on the ecosystem-scale model of Oikawa (1985), which has been used to simulate and analyze a tropical rain forest ecosystem (Oikawa, 1986), a temperate broad-leaved evergreen forest (Oikawa, 1998), and grassland (Oikawa, 1993; Alexandrov and Oikawa, 1995). In Sim-CYCLE, terrestrial carbon dynamics is conceptualized as a five-compartment system (Fig. 5.1). A water and radiation subscheme is required, because carbon dynamics is closely coupled with water and radiation budget (Fig. 5.1). Carbon in a given ecosystem ($WE$) is composed of plant biomass ($WP$) and soil organic carbon ($WS$). $WP$ is distributed in three compartments: foliage (subscript $F$), stem and branch (subscript $C$), and root (subscript $R$); $WS$ is distributed in two compartments: litter (subscript $L$) and mineral soil (subscript $H$):

$$WE = WP + WS \quad (5.1a)$$
$$WP = WP_F + WP_C + WP_R \quad (5.1b)$$
$$WS = WS_L + WS_H \quad (5.1c)$$

Atmosphere–biosphere CO$_2$ exchange occurs through three major processes: gross primary production ($GPP$), autotrophic plant respiration ($AR$), and heterotrophic soil respiration ($HR$). Net primary production ($NPP$), defined as:

$$NPP = GPP - AR \quad (5.2)$$
ecology but also for agronomy and forestry. Net ecosystem production ($NEP$), defined as follows:

$$NEP = NPP - HR$$ (5.3)

indicates the carbon balance of the ecosystem during a given period: a net sink or a source. Within a terrestrial ecosystem, carbon is transferred in the form of various organic compounds (largely carbohydrates) through several processes (Fig. 5.1), including translocation of photosynthate ($PT$), litterfall of dead biomass ($LF$), and synthesis of humic mineral soil ($HF$). At this stage, animal processes and lateral transportation of carbon from one ecosystem to another are neglected. The net change in each compartment ($A$) during a given period is given by the following equations:

$$\Delta WP_F = PT_F - ARG_F - LF_F$$ (5.4a)
$$\Delta WP_C = PT_C - ARG_C - LF_C$$ (5.4b)
$$\Delta WP_R = PT_R - ARG_R - LF_R$$ (5.4c)
$$\Delta WS_L = LF - HR_L - HF$$ (5.4d)
$$\Delta WS_H = HF - HR_H$$ (5.4e)

where photosynthate translocation ($PT$) is the difference between $GPP$ and $ARM$. It is apparent that $\Delta WE$ is equal to $NEP$. Since in Sim-CYCLE these fluxes are calculated monthly, the seasonal, interannual, and successional behaviors of terrestrial carbon dynamics can be simulated. By carrying out of these simulations, such important quantities as plant biomass, soil carbon storage, leaf area index ($LAI$), $NPP$, and $NEP$ can be estimated. The model calculation was launched at the juvenile stage, when initial carbon content was 0.1 Mg C ha$^{-1}$ for each compartment, and repeated for a period sufficiently long to attain the equilibrium state under stationary environmental conditions. The annual $NEP$ was used to determine whether the
ecosystem carbon dynamics had sufficiently equilibrated (i.e. climax stage): $NEP < 0.0001 \text{ Mg C ha}^{-1} \text{ year}^{-1}$.

5.2.1.2. Single-leaf processes

Single-leaf gas exchange is the most fundamental process in plant ecophysiology, which investigates the controls of various environmental factors on stomata, the leaf vents for photosynthesis and transpiration. The controls are described at a physiological scale and extended to a canopy scale (Fig. 5.2). The single-leaf photosynthetic rate ($PC$) is formulated as a Michaelis-type function of the incident $PPFD (PPFD_{IN})$:

$$PC = \frac{PC_{SAT} \cdot QE \cdot PPFD_{IN}}{PC_{SAT} + QE \cdot PPFD_{IN}}$$  \hspace{1cm} (5.5)$$

where $PC_{SAT}$ is the single-leaf photosynthetic rate under light-saturation, and $QE$ is light-use efficiency, or quantum yield of photosynthesis. $PC_{SAT}$ and $QE$ are complicated functions of temperature, $CO_2$ level, air humidity, and soil water, and they are different among biome types and between $C_3$ and $C_4$ species. They are formulated by using coefficient functions of important environmental factors, in a multiplicative way:

$$QE = QE_0 \cdot F_{QE}(TG) \cdot F_{QE}(CD_{ICL})$$  \hspace{1cm} (5.6)$$

$$PC_{SAT} = PC_{SAT0} \cdot F_{PC}(TG) \cdot F_{PC}(CD_{ICL}) \cdot F_{PC}(SW_L)$$  \hspace{1cm} (5.7)$$

where $QE_0$ and $PC_{SAT0}$ are potential maximum values under optimal conditions, and $F_{QE}$ and $F_{PC}$ denote the coefficient functions for temperature, $CO_2$, and water conditions, respectively. Before describing these coefficient functions, the author first explains the regulation of single-leaf gas exchange through stomatal conductance ($GS$). Sim-CYCLE adopts the
Semi-empirical model of stomatal conductance described by Ball et al. (1987) and modified by Leuning (1990):

\[ GS = x_1 + \frac{x_2 \cdot PC}{(CD_{ATM} - CD_{CMP})(1 + VPD/X_3)} \]  
(5.8)

where \(CD_{ATM}\) is the atmospheric CO\(_2\) concentration, \(VPD\) is the vapor pressure deficit, \(CD_{CMP}\) is the CO\(_2\) compensation point of photosynthesis, and \(x_1, x_2\) and \(x_3\) are biome-specific parameters. Eq. (5.8) suggests that \(GS\) decreases with increasing atmospheric CO\(_2\) concentration, which is the case according to many observations (Morison and Gifford, 1983; Field et al., 1995). Since gas exchange through the leaf cuticle is negligible, the leaf intercellular CO\(_2\) concentration (\(CD_{ICL}\)) is defined as follows:

\[ CD_{ICL} = CD_{ATM} - \frac{PC}{GS/1.56} \]  
(5.9)

where 1.56 is a factor to convert \(GS\) into CO\(_2\) conductance. \(CD_{ICL}\), rather than \(CD_{ATM}\), is directly related to photosynthetic capacity, although \(CD_{ICL}\) tends to change in parallel with the \(CD_{ATM}\).

As with the photosynthetic quantum yield \(QE\), Ehleringer and Björkman (1977) showed that C\(_3\) and C\(_4\) species respond to temperature and CO\(_2\) conditions in a disparate manner. Typically, the \(QE\) of C\(_4\) species is virtually insensitive to surrounding conditions (i.e. \(F_{QE}(TG) = F_{QE}(CD_{ICL}) = 1\)), because their CO\(_2\) condensation mechanism eliminates photorespiration. In contrast, the \(QE\) of C\(_3\) species depends strongly on temperature (\(TG\)) and CO\(_2\) level (\(CD_{ICL}\)):

\[ F_{QE}(TG) = \frac{52 - TG}{3.5 + 0.75(52 - TG)} \]  
(5.10a)

\[ F_{QE}(CD_{ICL}) = \frac{CD_{ICL}}{90 + 0.6CD_{ICL}} \]  
(5.10b)
Ehleringer et al. (1997) suggested that the difference in $Q_E$ between $C_3$ and $C_4$ species may be so important that it determines the geographical distribution of these two photosynthetic types. Moreover, it is fully recognized that $P_{C_{SAT}}$ of $C_4$ species is much higher than that of $C_3$ species, and that $C_4$ photosynthesis is vulnerable to low temperatures (Pearcy and Ehleringer, 1984). Sim-CYCLE incorporates such findings into the corresponding coefficient functions. The temperature dependence function of $P_{C_{SAT}}$ is a bell-shaped curve, formulated as follows (Raich et al., 1991):

$$F_{PC}(TG) = \frac{(TG - T_{MAX})(TG - T_{MIN})}{(TG - T_{MAX})(TG - T_{MIN}) - (TG - T_{OPT})^2}$$  \hspace{1cm} (5.11)

where $T_{MAX}$, $T_{MIN}$, and $T_{OPT}$ are, respectively, the maximum, minimum, and optimum temperatures for photosynthesis. For $C_3$ plants, $T_{OPT}$ is a function of intercellular CO$_2$ concentration:

$$T_{OPT} = T_{OPT0} + 0.01 CD_{ICL}$$  \hspace{1cm} (5.12)

where $T_{OPT0}$ is the minimum value of $T_{OPT}$ at very low $CD_{ICL}$. Eqs. (5.9) and (5.12) indicate that increases of atmospheric CO$_2$ and temperature may be interactive, making a prediction difficult even at the physiological scale, as has been demonstrated by a biochemical model study (Long, 1991). The CO$_2$ dependence of $P_{C_{SAT}}$ is expressed by a Michaelis-type function:

$$F_{PC}(CD_{ICL}) = \frac{CD_{ICL} - CD_{CMP}}{KM_{CD} + CD_{ICL}}$$  \hspace{1cm} (5.13)

where $KM_{CD}$ is a parameter of CO$_2$ sensitivity. The relationship between $CD_{ICL}$ and $P_{C_{SAT}}$ is strongly affected by temperature, which alters $Q_E$ and the CO$_2$ compensation point ($CD_{CMP}$). At
lower temperatures, an elevated CO₂ concentration would have a small fertilization effect, as observed in tundra species (Tissue and Oechel, 1987). CDₐₐₜₚₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜₜ¢

\[
\text{\textit{CD}_{\text{CMP}}} = \text{\textit{CD}_{\text{CMP}}}_0 \left[ 1 + \delta_1 (TG - 20) + \delta_2 (TG - 20)^2 \right]
\]  

(5.14)

where \( \text{\textit{CD}_{\text{CMP}}}_0 \) is the control value at 20 °C, and \( \delta_1 \) and \( \delta_2 \) are parameters of temperature sensitivity. Apparently, \( F_{\text{PC}} \left( \text{\textit{CD}_{\text{ICL}}} \right) \) represents the stomatal limitation of water stress on photosynthesis, because \( \text{\textit{CD}_{\text{ICL}}} \) is strongly regulated by stomatal conductance, which responds to air humidity. On the other hand, \( F_{\text{PC}} \left( \text{\textit{MS}_{\text{LW}}} \right) \) indicates the non-stomatal limitation of water stress, that is, the direct effect of soil-water availability on photosynthetic capacity:

\[
F_{\text{PC}} \left( \text{\textit{MS}_{\text{LW}}} \right) = \frac{\text{\textit{MS}_{\text{LW}}}}{\text{\textit{KM}_{\text{SW}}} + \text{\textit{MS}_{\text{LW}}}}
\]  

(5.15)

where \( \text{\textit{KM}_{\text{SW}}} \) is a parameter for soil-water availability. Since the components of single-leaf gas exchange, i.e. \( \text{\textit{GS}}, \text{\textit{CD}_{\text{ICL}}}, \text{\textit{PC}}, \) and \( \text{\textit{QE}} \), change interactively, a stationary state is numerically found by iterative calculations of Eqs. (5.5)–(5.9), (5.10a), (5.10b), (5.11)–(5.15).

### 5.2.1.2. Ecosystem-scale processes

1. Photosynthesis

\( \text{GPP} \) is the ultimate origin of all organic carbon, through which atmospheric CO₂ is fixed into dry matter. Indeed, the scaling procedure in terms of \( \text{GPP} \) is one of the characteristics of Sim-CYCLE, such that \( \text{GPP} \) is estimated by the dry-matter production theory established by Monsi and Saeki (1953). In our parameterization, it was assumed that environmental factors, such as temperature, CO₂, and water, were similar among all leaves. Monsi and Saeki (1953)
first formulated the downward attenuation of PAR irradiance due to mutual shading of leaves in a canopy with accumulating leaf area, as follows:

\[ PPFD_{\text{IN}} = PPFD_{\text{TOP}} \exp(-KA \cdot LAI_{\text{CMF}}) \]  

(5.16)

where \( PPFD_{\text{TOP}} \) denotes the photosynthetic photon flux density at a canopy top. It is apparent that PAR irradiance attenuates exponentially with the biome-specific coefficient \( KA \), such that a leaf underlying the cumulative leaf area index \( LAI_{\text{CMF}} \) receives light equivalent to \( PPFD_{\text{IN}} \). If a spherical distribution of leaf inclination is assumed, then \( KA \) is a function of solar height:

\[ KA = \frac{KA_0}{\sin(SE_{MD})} \]  

(5.17)

where \( KA_0 \) is the value for vertical incident radiation (biome-specific value), and \( SE_{MD} \) is the solar elevation at midday. In Sim-CYCLE, the single-sided leaf area index \( LAI \) is a prognostic variable, given by the following:

\[ LAI = 0.5SLA W_F \]  

(5.18)

where \( SLA \) is the specific leaf area. Based on the above assumption with respect to light attenuation, \( PPFD_{\text{IN}} \) in Eq. (5.5) can be substituted with Eq. (5.16), and integrate for the total leaf area index \( LAI \), in order to obtain the instantaneous \( GPP_{\text{INS}} \) rate.

\[ GPP_{\text{INS}} = \int_0^{LAI} PCDLAI \]

\[ = \frac{PC_{\text{SAT}}}{KA} \left[ \ln(QE + KA \cdot PPFD_{\text{TOP}}) - \ln(QE + KA \cdot PPFD_{\text{TOP}}) \times \exp(-KA \cdot LAI) \right] \]  

(5.19)
According to Kuroiwa (1966), the diurnal change in $PPFD_{TOP}$ is approximated well by a sine-square curve with the peak at midday, as follows:

$$PPFD_{TOP} = PPFD_{MD} \sin^2 \left( \frac{360 \cdot t}{DL} \right)$$  \hspace{1cm} (5.20)

where $t$ is time since sunrise, $DL$ is day length, and $PPFD_{MD}$ is the irradiance of $PPFD_{TOP}$ at midday. After substituting $PPFD_{TOP}$ with Eq. (5.20), Eq. (5.19) is integrated for $DL$ to obtain the daily $GPP_{DAY}$ rate:

$$GPP_{DAY} = \epsilon \int_0^{DL} \int_0^{LA} PCdLAdt$$

$$= \frac{2\epsilon PC_{SAT} DL}{KA} \ln\left(1 + \sqrt{1 + \frac{KA \cdot QE \cdot PPFD_{MD}}{PC_{SAT}}} \right) - \ln\left(1 + \sqrt{1 + \frac{KA \cdot QE \cdot PPFD_{MD} \exp(-KA \cdot LA)}{PC_{SAT}}} \right)$$  \hspace{1cm} (5.21)

The units is converted from $\mu$mol CO$_2$ m$^{-2}$ per day to Mg C ha$^{-1}$ per day by multiplying a unit conversion factor $\epsilon$ ($= 4.32 \times 10^4$) and the number of days in a month.

2. Respiration

Although autotrophic plant respiration ($AR$) is a major CO$_2$ flux comparable to $GPP$ and $NPP$ in magnitude, our ability to model $AR$ is insufficient, especially at an ecosystem scale. However, physiological studies (e.g. Amthor, 1989) suggest that $AR$ is composed of two components that have distinct functional meanings (i.e. for maintenance and for growth), and that different plant organs respire at different rates. $AR$ is calculated as a sum of six respiration rates, thus:
\[ AR = ARM + ARG \]

\[ AR = \sum_{x=\text{organ}}^{F,C,R} (ARM_x) + \sum_{x=\text{organ}}^{F,C,R} (ARG_x) \quad (5.22) \]

where \( ARM \) denotes the maintenance respiration of the whole plant, foliage, stem, and root, and similarly \( ARG \) denotes growth respiration. \( ARM \) is a function of the amount of existing carbon (\( WP \)) and temperature (\( TG \)):

\[ ARM_x = SARM_x \exp\left[ \frac{\ln(QT)}{10} (TG - 15) \right] WP_x \quad (5.23) \]

where \( SARM \) is the specific respiration rate, and \( TG \) is the temperature (control temperature, 15 °C). \( QT \) represents the sensitivity to temperature change. A typical value for \( QT \) is 2.0, and it ranges from 1.0 to 3.0 (Ryan, 1991) for various biomes and plant organs. However, as implied by Paembonan et al. (1992), Yokota and Hagihara (1996), \( QT \) may vary seasonally in temperate and boreal ecosystems that experience a large temperature change between summer and winter.

Based on their findings, \( QT \) is formulated as a function of temperature:

\[ QT = 2.0 \exp[ -0.009(TG - 15) ] \quad (5.24) \]

This modification results in a reduced sensitivity of \( ARM \) at higher temperatures, but very limited data are available to characterize biome specificity. In forest ecosystems, the maintenance respiration rate per unit biomass (\( SARM \)) for stem \( WP_c \) and root \( WP_r \) would decrease as passive woody tissues, or heartwood, accumulate (Yokota et al., 1994). The size-dependence of \( SARM \) can be approximated by a function of standing biomass, given as:

\[ SARM = \phi_1 + \frac{\phi_2}{\phi_3 + \exp(\phi_4(WP - \phi_5))} \quad (5.25) \]
where $\phi_1$ to $\phi_3$ are parameters for size-dependence. On the other hand, $ARG$ is not an explicit function of environmental factors but of plant growth rate, because $ARG$ represents the cost to produce new biomass (Amthor, 1989). Thus, $ARG$ is calculated only when biomass has a net gain (i.e. $\Delta WP > 0$), as follows:

$$
ARG_x = SARG_x \Delta WP_x = SARG_x \frac{PT_x}{1 + SARG_x}
$$

(5.26)

where $SARG$ is the specific growth respiration rate (foliage has a higher value than stem or root) and $PT$ is the photosynthate translocation to each organ (cf. Eq. (5.33)). Consequently, $ARG$ is indirectly regulated by environmental factors, via the $GPP$ and $ARM$ rates.

3. Decomposition

Soil organic carbon was divided into two compartments, because the decomposition rate differs greatly between them. The labile part of litter ($WSL$) circulates once every few months or years, while the passive part in mineral soil ($WSH$) lingers for decades or centuries. Heterotrophic soil respiration ($HR$) is composed of two constituents from each compartment:

$$
HR = HR_L + HR_H
$$

(5.27)

Both $HR_L$ and $HR_H$ are affected by temperature and soil moisture conditions:

$$
HR_L = SHR_L \cdot WSL \cdot F_{HR_L} (TS_{10}) \cdot F_{HR_L} (MS_{UP})
$$

(5.28a)

$$
HR_H = SHR_H \cdot WSH \cdot F_{HR_H} (TS_{200}) \cdot F_{HR_H} (MS_{LW})
$$

(5.28b)
where $SHR_t$ and $SHR_h$ are the specific respiration rates and $F_{HRt}$ and $F_{HRh}$ are the coefficient functions for temperature and soil-moisture conditions, respectively. For temperature dependence, an exponential function similar to Eq. (5.23) has been frequently used, but Lloyd and Taylor (1994) showed that it is not the best model, and they alternatively proposed an Arrhenius-type model:

$$F_{HR}(TS) = \exp\left[308.56\left(\frac{1}{56.02} - \frac{1}{TS + 46.02}\right)\right]$$ (5.29)

where $TS$ is soil temperature ($T_{S10}$ for $WS_L$ and $T_{S200}$ for $WS_H$). Eq. (5.29) resulted in a lower responsiveness at higher temperatures than does the exponential equation (Lloyd and Taylor, 1994). Although this process does not explicitly include the effect of soil freezing, it would have a negligible effect on $HR$ estimation. The coefficient function with respect to soil moisture $F_{HR}(MS)$ is the minimum of two contrasting components:

$$F_{HR}(MS) = \min\{F_{HR}(WA), F_{HR}(EA)\}$$ (5.30)

$F_{HR}(WA)$ represents the effect of soil moisture on microbial activity, while $F_{HR}(AE)$ represents the effect of soil air space (i.e. aerobic or anaerobic conditions):

$$F_{HR}(WA) = WA_0 + \frac{MS/WHC}{KM_{WA} + MS/WHC}$$ (5.31a)

$$F_{HR}(AE) = AE_0 + \frac{MS/WHC}{KM_{AE} + MS/WHC}$$ (5.31b)

where $WA_0$ and $AE_0$ are minimum values of, $KM_{WA}$ and $KM_{AE}$ are parameters related to responsiveness, and $WHC$ is water-holding capacity. In sum, a larger $MS$ increases $HR$ under dry conditions, whereas a larger $MS$ reduces $HR$ near the water-saturation point.
4. Litterfall

In the equilibrium state, the amount of annual LF must be identical to annual NPP, so plant biomass becomes stable. However, the shedding of dead biomass is one of the most difficult processes for mechanistic models to simulate. Consequently, a constant mortality or turnover rate irrespective of environmental conditions is assumed for evergreen biomes:

\[ LF = \sum_{x=\text{organ}}^{F,C,R} (SLF_x WP_x) \]  

(5.32)

where SLF is the specific litter-fall rate, or mortality. In grasslands, plants shed most of their shoots (WP_f and WP_e) during the winter (C₃, TG > 5 °C; C₄, TG >8 °C) or the dry season (MS > 0.1 x WHC), while the roots have a constant mortality.

5. Photosynthate translocation

Although a mechanistic model of photosynthate allocation (PT) is far from sufficient, the dry-matter production theory provides keys for addressing the problem. First, Monsi (1960) created a schematic diagram of plant growth, from CO₂ assimilation to biomass incrementation. This scheme suggests that assimilated carbon should be partitioned among plant organs, after subtracting the maintenance cost of the organs (ARM):

\[ EP = GPP - ARM = \sum_{x=\text{organ}}^{F,C,R} (PT_x) \]  

(5.33)

where EP is effective photosynthate for growth. If EP is negative, translocation and vegetative growth cannot be expected. Accordingly, a positive EP is partitioned among WP_f, WP_c, and WP_r, such that the fixed carbon is utilized by the plants most profitably to survive. Second,
Kuroiwa (1966) derived the optimal leaf area index \( \text{LAI}_{\text{OPT}} \) to maximize daily net carbon uptake from the daily \( \text{GPP} \) estimate using Eq. (5.21), as follows:

\[
\text{LAI}_{\text{OPT}} = \frac{1}{K_A} \ln \left[ \frac{K_A \cdot \text{QE} \cdot \text{PAR}_M}{\frac{\text{PC}_{\text{SAT}} \cdot \text{DL}}{\left( \text{PC}_{\text{SAT}} - \text{DCST} \right) - 1}} \right]
\]  

(5.34)

where \( \text{DCST} \) is the daily cost of maintaining a unit amount of foliage:

\[
\text{DCST} = \frac{\text{SARM}_F + \text{SLF}_F (1 + \text{SARG}_F)}{2\text{SLA}}
\]

(5.35)

When the community \( \text{LAI} \) is identical to the \( \text{LAI}_{\text{OPT}} \), almost all leaves, even those at the bottom of the canopy, may perform as a carbon source. In other words, this model prohibits the existence of heterotrophic leaves. Many empirical studies support shoot independence in terms of carbon economy; therefore, the theoretical prediction of \( \text{LAI}_{\text{OPT}} \) will be a valid indicator of carbon allocation to \( \text{WP}_F \). Thus, if \( \text{EP} \) is positive, photosynthate is allocated first to \( \text{WP}_F \) so that \( \text{LAI} \) becomes close to \( \text{LAI}_{\text{OPT}} \). Note that to avoid overshooting, an excessive amount of carbon should not be allocated to \( \text{WP}_F \), and that the \( \text{LAI}_{\text{OPT}} \) can be enlarged by stimulating the photosynthetic properties, i.e. \( \text{QE} \) and \( \text{PC}_{\text{SAT}} \). Residual carbon is allocated to \( \text{WP}_C \) and \( \text{WP}_R \) with a constant ratio, in order to realize the biome-specific growth form; woody biomes invest a considerable fraction to stem \( \text{WP}_C \), while in herbaceous biomes mostly is allocated to root \( \text{WP}_R \).

### 5.2.2. Soil thermal profile

When the soil water was freezing, the decline of the rate of thermal change induced by the latent heat (334.7 J g\(^{-1}\)), and the changes of the specific heat and thermal conductivity induced by the phase change of soil water vary the heat transfer in the soil layer. Especially at the temperature of near 0 °C, the former affects strongly. The vertical thermal profile of frozen
soil is calculated by changing the specific heat at the temperature of near 0 °C (Fukuda et al., 1980; Fig. 5.4).

The heat conduction equations in the frozen and unfrozen soil layer are described as follows:

\[
\frac{\Delta TS}{\Delta t} = K_f / C_f \rho \cdot \Delta^2 TS / \Delta Z^2 \quad (5.36a)
\]

\[
\frac{\Delta TS}{\Delta t} = K_u / C_u \rho \cdot \Delta^2 TS / \Delta Z^2 \quad (5.36b)
\]

where \(TS\) is the soil temperature (°C), \(Z\) is the depth (m), \(C_f \rho\) and \(C_u \rho\) are the specific heat of frozen and unfrozen soil (J m\(^{-3}\) K\(^{-1}\)), \(K_f\) and \(K_u\) are the thermal conductivities of frozen and unfrozen soil (W m\(^{-1}\) K\(^{-1}\)). The latent heat generation induced by soil water freezing reflects on increasing the specific heat near the freezing point, and the heat conductivity equation is rewritten as follows:

\[
\frac{\Delta TS}{\Delta t} = K_u / [MS \cdot L + 1/2 (C_u \rho + C_f \rho)] \cdot \Delta^2 TS / \Delta Z^2 \quad (5.37)
\]

where \(MS\) is the volumetric soil water content (m\(^3\) m\(^{-3}\)), \(L\) is the latent heat of freezing and thawing (344.7 J g\(^{-1}\)).

The three heat conduction equations described above are solved by the boundary condition, initial condition and the “Pure Implicit”-type difference equation as follows (Fig. 5.4):

\[
T(p + 1, n) = \theta (T(p + 1, n + 1) + T(p + 1, n - 1) - 2T(p + 1, n)) + T(p, n) \quad (5.38)
\]

\[
\theta = A \cdot \Delta t / (\Delta x)^2 \quad (5.39)
\]

where \(A\) is:

\[
K_f / C_f \rho \quad \text{ (T<=-1)} \quad (5.40a)
\]
Where $T(t,n)$ is the temperature at the time of $\Delta t$ (sec) x p and the location of $\Delta x \times n$ (m). The boundary condition is given for the top and bottom soil layer temperature, and the calculation is repeated to be at a steady state.

### 5.2.2.1. Thermal parameter

1. Specific heat

The volumetric specific heat $C$ of soil is defined as a sum of the specific heat of all soil components.

$$C = C_m \phi_m + C_w \theta + C_o \phi_o + C_o \phi_o$$  \hfill (5.41)

where $C$ is the volumetric specific heat (J m$^3$ K$^{-1}$), $\theta$ is the volumetric soil water content (cm$^3$ cm$^{-3}$), $\phi$ is volumetric fraction of each phase, subscripts $m$, $w$, $a$, and $o$ indicate mineral soil, water, air, and organic matter, respectively. $C_a$ is very small, and $C_o$ is similar to $C_m$. So the $C$ could be simplified as follows:

$$C = C_m (1 - \phi_f) + C_w \theta$$  \hfill (5.42)

where $\phi_f$ is the porosity, $C_m$ is adopted for 2.01, and $C_w$ is adopted for 4.18 (water) and 1.88 (ice) (de Vries, 1963).

2. Thermal conductivity

The thermal conductivity of soil $K$ depends on the solid density, water content, quartz density and organic matter content. Their qualitative relationship was examined in many
experimental studies. However, there is no universal equation to apply for various soil textures. This study adopted the experimental equation derived by Kersten (1963):

- for unfrozen sandy soil:

\[
K_u = 0.1442 \left(0.7 \cdot \log(\theta) + 0.4\right) \cdot 10^{0.6243} \cdot \rho_s
\]  
(5.43a)

- for frozen sandy soil:

\[
K_f = 0.01096 \cdot 10^{0.8116} \cdot \rho_s + 0.00461 \cdot \theta \cdot 10^{0.9115} \cdot \rho_s
\]  
(5.43b)

where \( \rho_s \) is the bulk density of soil (Mg m\(^{-3}\)).

### 5.2.3. Water budget

The water cycle in terrestrial ecosystems is simulated by a subscheme (Fig. 5.1) simply. The water storage is divided into three compartments, upper layer \((MS_{up}, \text{from } 0 \text{ to } 30 \text{ cm depth})\), lower layer \((MS_{lw}, \text{from } 30 \text{ cm to rooting depth})\), and snow accumulation \((SNA)\). For each water compartment, net balance of water content \((\Delta)\) during a given period is represented as follows:

\[
\Delta MS_{up} = PR_{\text{rain}} + TW - (EV + TR_{up}) - PN
\]  
(5.44)

\[
\Delta MS_{lw} = PN - TR_{lw} - RO
\]  
(5.45)

\[
\Delta SNA = PR_{\text{snow}} - TW
\]  
(5.46)

where \( MS \) is the soil water content (mm), \( SNA \) is the amount of snow cover (mm), \( PR_{\text{rain/snow}} \) are the rainfall and snowfall (mm), \( TW \) is thawing snow (mm), \( EV \) and \( TR \) are evaporation and transpiration (mm), \( PN \) is the penetration from upper storage to lower storage (mm), \( RO \) is the runoff (mm), subscript \( up \) and \( lw \) indicate the upper and lower storage, respectively. The calculations of each water budget component are explained in the following subsections.
5.2.3.1. Rainfall, snow and thawing

The amounts of the rainfall and snowfall in the total precipitation are estimated in the following equation:

\[ PR_{\text{rain}} = PR - PR_{\text{snow}} \]  \hspace{1cm} (5.47)
\[ PR_{\text{snow}} = PR/(1 + \exp(0.75 \cdot TA - 2.0)) \]  \hspace{1cm} (5.48)

where \( PR \) is the total precipitation (mm), \( TA \) is air temperature at a height of 2.0 m (°C). The snow thawing is estimated as a function of the soil temperature as follows:

\[ TW = SNA/(1 + \exp(-0.375 \cdot TG)) \]  \hspace{1cm} (5.49)

where \( TG \) is the soil surface temperature (°C).

5.2.3.2. Evaporation and transpiration

The S-W model also uses an electrical analogy to treat vertical water vapor movement in the SPAC circulation (Fig. 5.5), similar to the P-M model. The S-W model combines a one-dimensional model of crop transpiration and a one-dimensional model of soil evaporation. Surface resistances regulate the heat and mass transfer at the plant and soil surfaces, and aerodynamic resistances regulate those between the surfaces and the atmospheric boundary layer.

\[ \lambda ET = \lambda T + \lambda E = \lambda C_c \cdot T_o + \lambda C_s \cdot E_o \]  \hspace{1cm} (5.50)

where \( \lambda ET \) is the sum of the latent heat flux from the crop (\( \lambda T \)) and soil (\( \lambda E \)) (W m\(^{-2}\)), \( \lambda T_o \) and \( \lambda E_o \) are terms similar to those in the Penman-Monteith model (Monteith, 1965) that would apply to transpiration from the canopy and evaporation from the soil, respectively. \( C_c \) and \( C_s \) are the
canopy resistance coefficient and soil surface resistance coefficient, respectively. They are obtained as follows:

\[
\lambda T_o = \frac{\Delta R + (\rho_D D - \Delta r_{as} R_s) / (r_{aa} + r_{ac})}{\Delta + \gamma(1 + r_{sc} / (r_{aa} + r_{ac}))} \tag{5.51}
\]

\[
\lambda E_o = \frac{\Delta R + (\rho_D D - \Delta r_{as} (R - R_s)) / (r_{aa} + r_{as})}{\Delta + \gamma(1 + r_{ss} / (r_{aa} + r_{as}))} \tag{5.52}
\]

\[
C_c = \frac{1}{1 + \rho_c \rho_a / \rho_s (\rho_c + \rho_a)} \tag{5.53}
\]

\[
C_s = \frac{1}{1 + \rho_s \rho_a / \rho_c (\rho_s + \rho_a)} \tag{5.54}
\]

\[
\rho_a = (\Delta + \gamma)r_{aa} \tag{5.55}
\]

\[
\rho_c = (\Delta + \gamma)r_{ac} + \gamma r_{sc} \tag{5.56}
\]

\[
\rho_s = (\Delta + \gamma)r_{as} + \gamma r_{ss} \tag{5.57}
\]

where \(r_{sc}\) is the canopy resistance, and \(r_{ac}\) is the aerodynamic resistance of the canopy to in-canopy flow, and \(r_{ss}\) is the soil surface resistance. \(r_a\) and \(r_{as}\) are eddy diffusion resistances from the reference height to in-canopy heat exchange plane height and from there to the soil surface, respectively. \(R\) and \(R_s\) are the active radiation at the top of the canopy and the soil surface, respectively, and are defined as follows:

\[
R = R_n - G \tag{5.58}
\]

\[
R_s = R_{ns} - G \tag{5.59}
\]

where \(R_n\) and \(R_{ns}\) are net radiation fluxes into the complete canopy and the substrate (W m\(^{-2}\)), respectively, and \(G\) is soil heat flux (W m\(^{-2}\)).
The radiation reaching the soil surface can be calculated using Beer's law as follows (e.g. Ross, 1981).

\[ R_{rs} = R_s \exp(-CLAI) \]  

(5.60)

where \( C \) is the extinction coefficient of light attenuation, for which 0.7 was adopted (Monteith, 1973).

1. Stomatal and leaf boundary resistances

The canopy resistance \( r_{sc} \), and the aerodynamic resistance \( r_{ac} \) are defined as values dividing the bulk stomatal resistance \( r_{st} \) and the mean boundary layer resistance \( r_{b} \) of the canopy, respectively, by the sum of the abaxial and adaxial leaf areas.

\[ r_{sc} = r_{st} / 2LAI \]  

(5.61)

\[ r_{ac} = r_{b} / 2LAI \]  

(5.62)

where \( r_{st} \) was taken as an inverse number of stomatal conductance \( GS \). \( r_{b} \) was taken as a typical value of 25 s m\(^{-1}\) (Denmead, 1976; Uchijima, 1976).

2. Soil Surface Resistance

The soil surface resistance, \( r_{ss} \), is the resistance to water vapor movement from the interior to the surface of the soil, and was calculated using Kondo's model (Kondo et al., 1990).

\[ r_{ss} = F(\theta)/D_{am} \]  

(5.63)

\[ F(\theta) = F_1(\theta_{sat} - \theta)^{F_2} \]  

(5.64)

where \( \theta \) is the volumetric soil water content (cm\(^3\) cm\(^{-3}\)), \( T \) is the soil surface temperature (K). \( F(\theta) \) is the water vapor diffusion distance (m) as a function of \( \theta \) introduced by Kondo et
al.(1990) that developed the simple model of evaporation from the bare soil surface in laboratory experiments. In loamy soil, \( F_1 \) is \( 2.16 \times 10^2 \) (m) and \( F_2 \) is 10.0. \( \theta_{sat} \) is the saturated volumetric soil water content, and was taken as 0.54 m\(^3\) m\(^{-3}\) of the Mat Cry-gelic soil value at the experimental site (Cao et al., 1998). The molecular vapor diffusivity for movement from the interior to the surface of the soil, \( D_{am} \) has been given by Camillo et al. (1983)

\[
D_{am} = D_0 \left( \frac{T_s}{273.16} \right)^{1.75}
\]

(5.65)

where \( D_0 = 0.229 \times 10^{-4} \) (m\(^2\) s\(^{-1}\)) is the diffusivity of water vapor in soil air at the standard state; (Camillo et al., 1983).

3. Eddy Diffusion Resistance

The aerodynamic resistance of the mean canopy flow, \( r_{au} \), and the aerodynamic resistance of the soil surface to in-canopy flow, \( r_{as} \), were calculated from the vertical wind profile at the field and the eddy diffusion coefficient. In this paper, the scheme is referred to as Shuttleworth and Gurney (1990). This empirically and simply simulates the more important results of a second-order closure theory of in-canopy turbulence and, for this reason, is arguably superior to the earlier submodel in Shuttleworth and Wallace (1985), in both diagnostic and predictive applications of its theory. Above the canopy height, the eddy diffusion coefficient, \( K \), is given by

\[
K = k u_* (z - d)
\]

(5.66)

where \( u_* \), for neutral atmospheric stability, is given as follows:

\[
u_* = ku_*/\ln \left[ (z - d)/z_0 \right]
\]

(5.67)
where $u$ is the wind speed at the reference height $z$, $d$, $z_0$ is the zero plane displacement and the roughness length (m), and are given by the relational expression as a fixed fraction of crop height $h$ as presented by Monteith (1973):

\[
\begin{align*}
d &= 0.63h \\
z_0 &= 0.13h
\end{align*}
\]  

(5.68)  

(5.69)

Beneath the canopy height, the exponential decrease of the eddy diffusion coefficient, $K$, through the canopy, is given as follows:

\[
K = K_0 \exp\left[-n(1 - Z/h)\right]
\]  

(5.70)

where $K_0$ is the eddy diffusion coefficient at the top of canopy, and $n$ is the extinction coefficient of the eddy diffusion, chosen as a typical value for crops (wheat, rice, clover and corn), 2.5 (Monteith, 1973). $K_0$ is determined as follows:

\[
K_0 = k_u (h - d)
\]  

(5.71)

$r_{as}$ and $r_{ao}$ are assumed as integrations of Eqs. (5.66) and (5.70), respectively, over the height ranges $0$ to $(z_0 + d)$ and $(z_0 + d)$ to $z$; thus

\[
\begin{align*}
r_{as} &= \frac{h \exp(n)}{n K_h} \left[\exp(-n z_0'/h) - \exp\left[-n (z_0 + d)/h\right]\right] \\
r_{ao} &= \frac{1}{k u_p} \left[\ln\left(\frac{z - d}{h - d}\right) + \frac{h}{n K_h} \left[\exp\left\{n \left(1 - (z_0 + d)/h\right)\right\} - 1\right]\right]
\end{align*}
\]  

(5.72)  

(5.73)

where $z_0' = 0.01$ m (van Bavel and Hillel, 1976) is the roughness length of the substrate.
5.2.3.3. Runoff and Infiltration

Two buckets are settled for runoff and infiltration calculations (Fig. 5.6). The overflow from the upper bucket comes into the lower bucket as an infiltration. The overflow from the lower bucket outflows from the ecosystem as a runoff. In a new bucket model (Kondo, 1993), the soil water content $MS$, the inflow water into the ground $RI$, and the runoff $RO$ are expressed in the following equations.

(a) in the upper bucket

\[ \Delta MS_{up} = (MS_{up_{max}} - MS_{up}) \cdot \tanh(X_{up}) \]  \hspace{1cm} (5.74)

where $MS_{up_{max}}$ is the maximum water storage (mm), and $MS_{up}$ is the water storage in the previous time step (mm). $X_{up}$ and $\tanh(X_{up})$ are described as follows:

\[ X_{up} = RI/(MS_{up_{max}} - MS_{up}) \]  \hspace{1cm} (5.75)

\[ RI = PR_{rain} + TW - (EV + TR_{up}) \]  \hspace{1cm} (5.76)

\[ \tanh(X_{up}) = (1 - \exp(-2X_{up}))/\{1 + \exp(-2X_{up})\} \]  \hspace{1cm} (5.77)

From these equations above, the penetration from the upper bucket to the lower bucket $PN$ is defined as follows:

\[ PN = RI - \Delta MS_{up} \quad \text{\{PN > 0\}} \]  \hspace{1cm} (5.78)

(b) in the lower bucket

\[ \Delta MS_{lw} = (MS_{lw_{max}} - MS_{lw}) \cdot \tanh(X_{lw}) \]  \hspace{1cm} (5.79)
where \( MS_{bw_{\text{max}}} \) is the maximum water storage (mm), \( MS_{bw} \) is the water storage in the previous time step (mm). \( X_{bw} \) is described as follows:

\[
X_{bw} = \frac{PN}{(MS_{bw_{\text{max}}} - MS_{bw})}
\]

(5.80)

From these equations above, the runoff from the lower bucket \( RO \) is defined as follows:

\[
RO = PN - \Delta MS_{bw} \quad \text{ (} RO > 0 \text{ )}
\]

(5.81)

In the frozen soil, the moisture movement induced by unfrozen soil water should be taken into account. However, its hydraulic conductivity is as very small as in the order of \( 10^{-8} \) (m s\(^{-1}\)) at the temperature of 0 °C, in the order of \( 10^{-13} \) (m s\(^{-1}\)) at the temperature of -1 °C and extremely smaller at the lower temperature. Consequently, the moisture movement in the frozen soil is assumed to be negligible in this study.

### 5.3. Model experiment designs

Three model experiments were made in this study: a steady state simulation, a transient simulation and a sensitivity analysis. All experiments used the monthly climate data measured at the meteorological observation filed of the Haibei station, the Chinese Academy of the Sciences for twenty years from 1981 to 2000 as the input climate data, and adopted the ecophysiological parameters (Fig. 5.3, Table 5.1) equipped as a Tibetan meadow ecosystem (No. 20 biome) in the Sim-CYCLE.

In a steady state simulation, calculation is launched from the juvenile stage, where initial carbon content is 0.1 Mg C ha\(^{-1}\) for each compartment, and repeated for a sufficiently long period to attain the equilibrium state under a stationary environmental condition using 20 years averaged monthly climate condition for 1981-2000 (Fig. 5.7). Annual \( NEP \) is used as the criteria to determine whether the ecosystem carbon dynamics is sufficiently equilibrated (i.e. climax
stage): \(NEP < 0.0001 \text{ Mg C ha}^{-1} \text{ yr}^{-1}\). The resulting equilibrated carbon dynamics was compared to observed values for examination of the model estimation accuracy.

In a transient simulation, calculation is started from an equilibrium state, where annual \(NEP\) is less than 0.0001 Mg C ha\(^{-1}\) yr\(^{-1}\) after the steady state simulation. The carbon dynamics was simulated for 1981-2000 using the 20 years transient monthly climate data, and compared with the climate perturbations and global climatic events, e.g. volcanic eruptions and ENSO.

In a sensitivity analysis, calculation is also started from the equilibrium state where annual \(NEP\) is less than 0.0001 Mg C ha\(^{-1}\) yr\(^{-1}\) after the steady state simulation. The carbon dynamics was simulated for the equilibrium state using the 20 years averaged monthly climate data including the prescribed climate shifts, e.g. \(\pm 10 \degree C\) in air temperature, \(\pm 50\%\) in precipitation and 200-1000 ppm in CO\(_2\) concentration.

5.4. Results

5.4.1. Steady state experiment

In the steady state experiment using 20 years, 1981-2000, averaged monthly climate data, 1003 years are necessary to equilibrate the annual \(NEP\). Figure 5.8 shows the growth of ecosystem carbon fluxes and storages for the first 200 years. \(GPP, AR\) and \(NPP\) increased in the beginning of calculations rapidly, and were re-equilibrated in the 50 years. On the other hand, the \(HR\) increased slowly, and \(NEP = (NPP-HR)\) decreased slowly toward zero. The plant biomass \(W_{plant}\) and \(LAI\), gas exchange area in the photosynthesis, increased in the beginning of calculations rapidly, and were re-equilibrated in the 50 years. The soil microbes and litter biomass \(W_{soil}\), determined by the input of plant litterfall, increased slowly.

The annual carbon fluxes, equilibrated by Sim-CYCLE, were compared with the observed and estimated values around the Haibei station derived from the literatures (Fig. 5.9; Table. 5.2). Although the \(GPPs\) are 1.5 times larger than the observed values (Sim-CYCLE, 847.8; Observation, 575.1-527.4), the other fluxes are as much as the observed values (\(NPP; \) Sim-CYCLE, 427.8; Observation, 124.2-823.0; \(SR\); Sim-CYCLE, 602.3; Observation, 578.0-2721.0).
In the seasonal changes of the carbon fluxes, the GPP, AR and NPP are positive from May to September, and zero during other months (Fig. 5.10). The HR is positive from May to November, and zero during other months (Fig. 5.10). The NEP reached maximum state in July and August, and decreased to be negative in October. In the seasonal changes of the carbon storages, the leaf, and stem biomass and LAI are positive from May to September (Fig. 5.11). The root biomass increases from May to October gradually, and the litter biomass increases rapidly in October induced by the plant litterfall. The mineral soil biomass does not change during whole year.

The soil thermal profiles show a good agreement with the measured data except for June and July (Fig. 5.12). The soil freezing depths are also estimated successfully. The heat budgets are shown in Fig. 5.13. The Rn reached its maximum in June and the G reached its maximum in April and minimum in November. The H reached its maximum in May and the λE reached maximum in July. The water budgets are shown in Fig. 5.14. The TR reached its maximum status in summer (July and August), when the LAI reached the maximum status (Fig. 5.11). The EV was large in spring (April and May) and the end of autumn (October), when the LAI was low (Fig. 5.11). The RO existed during the growing season. The TW reached its maximum in the spring (April). The SWC\textsubscript{w} started to increase from May and reached its maximum in September, and the SWC\textsubscript{w} did not change significantly.

5.4.2. Transient experiment

In the transient state experiment using the 20 years, transient climate data, the averaged annual carbon fluxes were similar to the observed and estimated values around the Haibei station derived from the literatures (Table 5.2). Aboveground NPPs were slightly larger than the observed values for 1981-1993 (Fig. 5.16).

In the interannual changes of the carbon fluxes and storages (Fig. 5.16), the GPPs ranged from 7.66 to 9.11, the ARs from 3.95 to 4.64, the NPPs from 3.71 to 4.47, the HRs from 3.61 to 4.41 and the NEPs from −0.65 to +0.65 (Mg C ha\textsuperscript{-2} yr\textsuperscript{-1}). The \( W_{\text{plant}} \) ranged from 6.63 to
6.75. The $W_{soil}$ reached a maximum of 207.25 (Mg C ha$^{-2}$) in 1984 and decreased to 205.82 (Mg C ha$^{-2}$) in 2000 (Table 5.3).

In the comparisons of the climate and carbon flux anomalies, the variation from their 20 years averaged values (Fig. 5.17; Table 5.4), the GPP, AR, NPP, NEP anomalies are synchronized with each other. The GPP and NEP anomalies ranged in ±0.76 and ±0.68 Mg C ha$^{-2}$, respectively. The HR anomaly changed along with the surface temperature variations.

Carbon flux anomalies plotted against the climate (annual average temperature, annual precipitation and annual averaged solar radiation) anomalies are shown in Figs. 5.18, 19, and 20. The increment of annual averaged temperature increased $\Delta$GPP, $\Delta$AR and $\Delta$NPP slightly, but increased $\Delta$HR largely with relatively higher determinant coefficient, and decreased the resulting $\Delta$NEP (Fig. 5.18). The increment of annual precipitation and annual averaged solar radiation increased or decreased the carbon fluxes so slightly with relatively lower determinant coefficient (Figs. 5.19 and 5.20).

Figure 5.22 shows the seasonal changes in the slope of regression lines between temperature, precipitation, solar radiation and carbon flux anomalies. The increment of temperature in June to August increased the HR and decreased the NEP, and the increment of temperature in September increased the GPP, AR and NPP. The fluctuations of precipitation seemed not to affect the carbon fluxes. The increment of solar radiation from June to August increased the GPP, AR and NPP.

5.4.3. Model sensitivity analysis

In the model sensitivity analysis, the time courses of carbon fluxes and storage variation, calculated using 20 years averaged climate data with the increment of temperature of 5 °C, solar radiation of 10 % and precipitation of 30 % and the CO₂ concentration of 700 ppm, were shown in Figs. 5.23 and 5.24. These results show the potential responses of the carbon dynamics to climate changes. The GPP, AR and NPP increased rapidly in the beginning of the experiment, and were equilibrated after 20 years elapsed. In contrast, The HR increased slowly, and the NEP decreased slowly toward zero. The plant biomass $W_{plant}$ increased in the beginning of
calculations rapidly, and were re-equilibrated in the 20 years. The soil microbes and litter biomass $W_{soil}$ increased slowly.

Next, the equilibrated carbon, water and heat fluxes and storages calculated using 20 years averaged climate data with the temperature data of $\pm 10^\circ C$, the solar radiation data of $\pm 30\%$ and the precipitation data of $\pm 50\%$ and the CO$_2$ concentration data of 200-1000 ppm respectively, were plotted against the prescribed climate fluctuations in Fig. 5.25 and 5.26. The increment of temperature by the $5^\circ C$ increased the carbon fluxes and storages larger than the present values, and the increment over $7.5^\circ C$ decreased them reversely. The magnitude of photosynthesis (GPP) links to the magnitude of the transpiration (TR). The increment of temperature by the $3^\circ C$ increased the latent heat flux larger than the present values, and the increment over $5^\circ C$ did not increase them moreover. The increment of solar radiation and precipitation did not affect the carbon dynamics largely. The increment of solar radiation leads to increase the TR and decrease the soil water content significantly. The decrescent of precipitation leads to decrease the soil water content significantly. The increment of CO$_2$ concentration by the 700 ppm increased the carbon fluxes and storages, but that at the concentration of 1000 ppm did not change the carbon dynamics significantly.

### 5.5. Discussion

#### 5.5.1. Model accuracy

In the validation of the simulated annual carbon dynamics (Fig. 5.9; Table 5.2), the model estimated values coincided with the observed values, and this confirms the successful model simulation.

In the seasonal changes, the GPP and $R_s$ ($AR+HR$) coincided with the measured data by the eddy covariance method in 2002 (Fig. 5.10; Table 4.2). But, the NEP did not synchronized in the maximum and minimum timings with the measured data. However, it is not enough to judge whether this results are reasonable or not from the current limited measurement. Furthermore multi-year data acquisitions of the CO$_2$ exchanges are therefore required.
In the transient state experiment (Fig. 5.16), the simulated aboveground NPPs were slightly larger than the observed values for 1981-1993 (RMSE = 0.089 Mg ha\(^{-2}\); Fig. 5.15). Because the observed values may contain the sampling error, these relatively high agreements confirm the high model accuracy.

### 5.5.2. The Relationship between climate perturbations and ecosystem CO₂ exchanges

The increment of temperature increased the HR, and decreased the NEP reversely in the transient state experiment. This was caused by the microbial decomposition rate increasing in the mineral soil and litter biomass. On the other hand, there was no clear relationship between the precipitation fluctuations and carbon fluxes. This indicated that the water is not a limiting factor in this ecosystem in the contrast to other relatively dry grassland, e.g. steppes, savannas. There was also no clear relationship between the solar radiation fluctuations and carbon fluxes. This may be caused by the seasonality of plant response to solar radiation increment; the increment of solar radiation only from June to August increased the GPP, AR and VPP (Fig. 5.22).

The global climate events affected the ecosystem carbon dynamics. Mt. El Chichon eruption in 1982 lead to lower temperature and the decrements of GPP and Rₑ (AR+HR) in the next year 1983. The ENSO (El-Nino and Southern Oscillation) event in 1997 winter to 1998 summer lead to higher temperature and the increments of GPP and Rₑ (AR+HR) in 1998. Thus, the increments of temperature increased the ecosystem carbon fluxes. In contrast, Mt. Pinatubo eruption in 1991 lead to lower temperature and the increments of GPP and Rₑ (AR+HR) in the next year 1992. This increment may be caused by the increments of solar radiation in the summer of 1998 (Figs. 5.21 and 5.22). The anomalies in the NEPs did not synchronize with those in a global scale as presented by Ito and Oikawa (2000a). Resultingly, the NEPs tended to be negative and the \( W_{soil} \) decreased gradually for 20 years period. However it cannot be concluded that study site was a source of atmospheric CO₂, because the transient calculation in the carbon dynamics started from the neutral condition (i.e. \( NEP = 0 \)) and have the possibility to
balance out the long-term effects on ecosystem carbon uptake by global warming and CO2 fertilization.

5.5.3. The potential response of the ecosystem CO2 exchanges against global warming

The model sensitivity analysis showed that the GPP, AR, NPP and $W_{plant}$ responded quickly and the HR and $W_{soil}$ responded very slowly against the climate change. Consequently, the NEP was positive in the beginning of the model simulation. These may indicate that the long-term global warming will increase the ecosystem carbon uptake, and now the alpine meadow ecosystems maybe in this situation.

The increments of temperature by 5 °C increased the GPP and that over 7.5 °C decreased the GPP (Fig. 5.25). Zhang and Welker (1996) showed that while the peak community biomass showed no significant change, the duration of peak biomass was extended in the warming experiment by 5 °C of air temperature at the Haibei alpine meadow. Thus, the warming by 5 °C advances the ecosystem photosynthetic activity and increases the duration of growing season, and thus, the GPP. But, the warming over 7.5 °C may go beyond the optimum temperature of photosynthesis and decrease the GPP. In this extreme warming, this ecosystem may shift to another biome type that can survive the new climate.

The increments of CO2 concentration by 700 ppm increased the GPP, but that at the concentration of 1000 ppm carbon dynamics showed no significant change (Fig. 5.25). One possibility is that such high CO2 concentration as 1000 ppm induced the stomatal conductance.
<table>
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<th>Unit</th>
<th>Explanation</th>
<th>Remarks</th>
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Table 5.2. Comparisons of ecosystem carbon dynamics between Sim-CYCLE estimates and those from other studies

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135
Table 5.2. Continued

| Belowground NPP | Sim-CYCLE 215.1 (140.7**) gC m⁻² yr⁻¹ 1981-2000 steady state this study |
| Sim-CYCLE 210.9 ± 3.6 (139.0 ± 8.2**) gC m⁻² yr⁻¹ 1981-2000 transient state this study |
| Estimated from literature 274.3 gC m⁻² yr⁻¹ --- |

| NEP | Sim-CYCLE 0.0 gC m⁻² yr⁻¹ 1981-2000 steady state this study |
| Sim-CYCLE -3.6 ± 32.2 gC m⁻² yr⁻¹ 1981-2000 transient state this study |
| Eddy covariance 78.5 gC m⁻² yr⁻¹ 2002 gap-filled by the non-linear regression method this study |
| Eddy covariance 113.9 gC m⁻² yr⁻¹ 2002 gap-filled by the non-linear regression method this study |
| Eddy covariance 145.9 ± 3.3 gC m⁻² yr⁻¹ 2002 gap-filled by the MDV method this study |
| Eddy covariance 118.8 ± 18.9 gC m⁻² yr⁻¹ 2002 gap-filled by the LookUp method this study |

| Soil Respiration | Sim-CYCLE 602.3 gC m⁻² yr⁻¹ 1981-2000 steady state this study |
| Sim-CYCLE 596.3 ± 23.0 gC m⁻² yr⁻¹ 1981-2000 transient state this study |
| Observation 2721.0 gC m⁻² yr⁻¹ 1981-1993 alkali absorption method this study |
| Observation 578.0 gC m⁻² yr⁻¹ 1998-1999 closed chamber method this study |
| Observation 707.2 * gC m⁻² yr⁻¹ 1998-1999 closed chamber method this study |

* Aboveground NPP as a maximum aboveground biomass, ** Belowground NPP as a maximum belowground biomass
<table>
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<tr>
<th>Term</th>
<th>Method</th>
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<th>Unit</th>
<th>Year</th>
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<td>90.5 ± 3.5 (83.9 ± 3.2*)</td>
<td>gC m⁻²</td>
<td>1981-2000</td>
<td>transient state</td>
<td>this study</td>
</tr>
<tr>
<td></td>
<td>Estimated from literature</td>
<td>165.5</td>
<td>gC m⁻²</td>
<td>2001</td>
<td>mowed on 20 Aug.</td>
<td>Kato et al. (in press, 2004b)</td>
</tr>
<tr>
<td></td>
<td>Observation</td>
<td>127.3**</td>
<td>gC m⁻²</td>
<td>2002</td>
<td>Mowing</td>
<td>Kato et al. (in press, 2004b)</td>
</tr>
<tr>
<td></td>
<td>Estimated from literature</td>
<td>114.1</td>
<td>gC m⁻²</td>
<td>---</td>
<td></td>
<td>Luo et al. (2002)</td>
</tr>
<tr>
<td>Belowground biomass (root)</td>
<td>Sim-CYCLE</td>
<td>635.6 ± 45.4</td>
<td>gC m⁻²</td>
<td>1981-2000</td>
<td>steady state</td>
<td>this study</td>
</tr>
<tr>
<td></td>
<td>Observation</td>
<td>623.2 ± 44.6</td>
<td>gC m⁻²</td>
<td>1981-2000</td>
<td>transient state</td>
<td>this study</td>
</tr>
<tr>
<td></td>
<td>Estimated from literature</td>
<td>738.8</td>
<td>gC m⁻²</td>
<td>2001</td>
<td>mowed on 20 Aug.</td>
<td>Kato et al. (in press, 2004b)</td>
</tr>
<tr>
<td></td>
<td>Estimated from literature</td>
<td>903.5</td>
<td>gC m⁻²</td>
<td>---</td>
<td></td>
<td>Luo et al. (2002)</td>
</tr>
<tr>
<td>Litter biomass</td>
<td>Sim-CYCLE</td>
<td>1075.9 ± 26.3</td>
<td>gC m⁻²</td>
<td>1981-2000</td>
<td>steady state</td>
<td>this study</td>
</tr>
<tr>
<td></td>
<td>Sim-CYCLE</td>
<td>1074.1 ± 35.3</td>
<td>gC m⁻²</td>
<td>1981-2000</td>
<td>transient state</td>
<td>this study</td>
</tr>
<tr>
<td>Mineral soil</td>
<td>Sim-CYCLE</td>
<td>19586.9 ± 6.1</td>
<td>gC m⁻²</td>
<td>1981-2000</td>
<td>steady state</td>
<td>this study</td>
</tr>
<tr>
<td></td>
<td>Sim-CYCLE</td>
<td>19578.9 ± 34.3</td>
<td>gC m⁻²</td>
<td>1981-2000</td>
<td>transient state</td>
<td>this study</td>
</tr>
<tr>
<td></td>
<td>Estimated from literature</td>
<td>18200</td>
<td>gC m⁻²</td>
<td>---</td>
<td>assigned to alpine meadow grassland type</td>
<td>Ni, J. (2002)</td>
</tr>
<tr>
<td></td>
<td>Estimated from literature</td>
<td>36500</td>
<td>gC m⁻²</td>
<td>---</td>
<td>assigned to alpine meadow soil type</td>
<td>Fang et al. (1996)</td>
</tr>
<tr>
<td>Maximum Leaf Area Index</td>
<td>Sim-CYCLE</td>
<td>2.25</td>
<td>cm³ cm⁻³</td>
<td>1981-2000</td>
<td>steady state</td>
<td>this study</td>
</tr>
<tr>
<td></td>
<td>Sim-CYCLE</td>
<td>2.21 ± 0.08</td>
<td>cm³ cm⁻³</td>
<td>1981-2000</td>
<td>transient state</td>
<td>this study</td>
</tr>
<tr>
<td></td>
<td>Observation</td>
<td>3.08</td>
<td>cm³ cm⁻³</td>
<td>2001</td>
<td>mowed on 20 Aug.</td>
<td>Kato et al. (in press, 2004b)</td>
</tr>
<tr>
<td></td>
<td>Observation</td>
<td>2.70**</td>
<td>cm³ cm⁻³</td>
<td>2002</td>
<td>Mowing</td>
<td>Kato et al. (in press, 2004b)</td>
</tr>
</tbody>
</table>

Data shows annual values for single year estimations and measurements, annual averages and standard deviation for multiple years estimations and measurements (average ± s.d.). * maximum foliage biomass,  ** monthly averaged value from Kato et al. (in press, 2004b)
Table 5.4. Climate and carbon fluxes annual anomalies in the Haibe alpine meadow.

<table>
<thead>
<tr>
<th>Year</th>
<th>$\Delta T_{	ext{surface}}$ °C</th>
<th>$\Delta PR$ mm</th>
<th>$\Delta SWR$ W m$^{-2}$</th>
<th>$\Delta GPP$ Mg C ha$^{-1}$</th>
<th>$\Delta AR$ Mg C ha$^{-1}$</th>
<th>$\Delta NPP$ Mg C ha$^{-1}$</th>
<th>$\Delta HR$ Mg C ha$^{-1}$</th>
<th>$\Delta NEP$ Mg C ha$^{-1}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>1981</td>
<td>-0.43</td>
<td>-61.32</td>
<td>0.36</td>
<td>-0.30</td>
<td>-0.14</td>
<td>-0.17</td>
<td>0.09</td>
<td>-0.25</td>
</tr>
<tr>
<td>1982</td>
<td>-0.41</td>
<td>-105.62</td>
<td>1.75</td>
<td>0.35</td>
<td>0.16</td>
<td>0.19</td>
<td>-0.49</td>
<td>0.68</td>
</tr>
<tr>
<td>1983</td>
<td>-0.98</td>
<td>-31.52</td>
<td>1.73</td>
<td>-0.29</td>
<td>-0.16</td>
<td>-0.12</td>
<td>-0.30</td>
<td>0.17</td>
</tr>
<tr>
<td>1984</td>
<td>-0.72</td>
<td>-75.02</td>
<td>0.97</td>
<td>-0.28</td>
<td>-0.16</td>
<td>-0.11</td>
<td>-0.13</td>
<td>0.01</td>
</tr>
<tr>
<td>1985</td>
<td>-0.23</td>
<td>263.18</td>
<td>0.36</td>
<td>-0.69</td>
<td>-0.34</td>
<td>-0.35</td>
<td>0.12</td>
<td>-0.47</td>
</tr>
<tr>
<td>1986</td>
<td>-0.18</td>
<td>112.88</td>
<td>0.39</td>
<td>0.02</td>
<td>-0.01</td>
<td>0.03</td>
<td>-0.17</td>
<td>0.20</td>
</tr>
<tr>
<td>1987</td>
<td>0.47</td>
<td>57.98</td>
<td>-1.15</td>
<td>-0.16</td>
<td>-0.07</td>
<td>-0.08</td>
<td>-0.09</td>
<td>-0.09</td>
</tr>
<tr>
<td>1988</td>
<td>0.43</td>
<td>211.78</td>
<td>-2.27</td>
<td>-0.18</td>
<td>-0.08</td>
<td>-0.10</td>
<td>-0.09</td>
<td>-0.01</td>
</tr>
<tr>
<td>1989</td>
<td>0.38</td>
<td>289.08</td>
<td>-1.27</td>
<td>-0.23</td>
<td>-0.10</td>
<td>-0.12</td>
<td>0.03</td>
<td>-0.15</td>
</tr>
<tr>
<td>1990</td>
<td>0.44</td>
<td>-39.62</td>
<td>4.75</td>
<td>0.36</td>
<td>0.17</td>
<td>0.19</td>
<td>0.02</td>
<td>0.17</td>
</tr>
<tr>
<td>1991</td>
<td>-0.47</td>
<td>-136.02</td>
<td>23.34</td>
<td>-0.41</td>
<td>-0.19</td>
<td>-0.21</td>
<td>0.02</td>
<td>-0.23</td>
</tr>
<tr>
<td>1992</td>
<td>-0.82</td>
<td>1.38</td>
<td>44.96</td>
<td>0.12</td>
<td>0.03</td>
<td>0.09</td>
<td>-0.13</td>
<td>0.21</td>
</tr>
<tr>
<td>1993</td>
<td>0.04</td>
<td>-53.72</td>
<td>49.65</td>
<td>-0.07</td>
<td>-0.04</td>
<td>-0.03</td>
<td>0.11</td>
<td>-0.14</td>
</tr>
<tr>
<td>1994</td>
<td>0.85</td>
<td>-39.02</td>
<td>39.02</td>
<td>-0.55</td>
<td>-0.25</td>
<td>-0.31</td>
<td>0.31</td>
<td>-0.62</td>
</tr>
<tr>
<td>1995</td>
<td>-0.16</td>
<td>-8.42</td>
<td>39.02</td>
<td>0.76</td>
<td>0.35</td>
<td>0.41</td>
<td>-0.17</td>
<td>0.58</td>
</tr>
<tr>
<td>1996</td>
<td>-0.60</td>
<td>-55.92</td>
<td>-16.86</td>
<td>0.50</td>
<td>0.22</td>
<td>0.28</td>
<td>0.03</td>
<td>0.25</td>
</tr>
<tr>
<td>1997</td>
<td>0.12</td>
<td>-97.32</td>
<td>-47.27</td>
<td>-0.72</td>
<td>-0.34</td>
<td>-0.37</td>
<td>0.01</td>
<td>-0.38</td>
</tr>
<tr>
<td>1998</td>
<td>0.92</td>
<td>-19.92</td>
<td>-37.51</td>
<td>0.46</td>
<td>0.26</td>
<td>0.21</td>
<td>0.28</td>
<td>-0.08</td>
</tr>
<tr>
<td>1999</td>
<td>0.83</td>
<td>-153.62</td>
<td>-35.36</td>
<td>0.69</td>
<td>0.38</td>
<td>0.32</td>
<td>0.12</td>
<td>0.20</td>
</tr>
<tr>
<td>2000</td>
<td>0.51</td>
<td>-59.22</td>
<td>-37.07</td>
<td>0.59</td>
<td>0.32</td>
<td>0.27</td>
<td>0.32</td>
<td>-0.04</td>
</tr>
<tr>
<td>s.d.</td>
<td>0.58</td>
<td>125.65</td>
<td>26.43</td>
<td>0.46</td>
<td>0.23</td>
<td>0.23</td>
<td>0.20</td>
<td>0.32</td>
</tr>
</tbody>
</table>
Figure 5.1. Schematic diagram of Sim-CYCLE.
Figure 5.2. Schematic diagram of single leaf processes in Sim-CYCLE (Adopted from Ito, 2000). CO2 and H2O (vapor) are exchanged through stomata, i.e. photosynthesis $PC$ and transpiration $TR$, respectively. The conductance of stomata $GS$ is regulated by atmospheric CO2 concentration $CD_{ATM}$, vapor pressure deficit $VPD$, and photosynthetic rate $PC$. The term $PC$ is a function of irradiance of photosynthetically active radiation $PPFDIN$, intercellular CO2 concentration $CD_{ICL}$, surface temperature $TG$, and soil moisture content $MS$. In the calculation, photosynthetic properties, i.e. light use efficiency $LUE$, CO2 compensation point $CD_{CMP}$, and optimum temperature $T_{OPT}$, are also estimated. The term $TR$ is affected by net radiation $RNP$ and $VPD$ (a function of air vapor pressure $VP$ and air temperature $TA$).
Figure 5.3. Schematic diagram of the Sim-CYCLE calculation processes.
\[ T_{y,i+1} = rT_{y-1,i} + (1 - 2r)T_{y,i} + rT_{y+1,i} \]

\[ T_{y,i}, \text{ temperature at the horizon } y \text{ in the time step } i \text{ (y=1\sim32)} \]

\[ r, \text{ coefficient} \]

Figure 5.4. Schematic diagram of soil thermal profile processes in Sim-CYCLE.
Figure 5.5. Schematic diagram of a one-dimensional description of energy partitioning for a canopy. Illustration shows the S-W model description (Shuttleworth and Wallace, 1985). $\lambda ET$ is the evapotranspiration from the crop ($\lambda T$) and soil ($\lambda E$). $r_c$ and $r_{sc}$ are the canopy resistances, and $r_{ss}$ is the soil surface resistance. $r_a$ is the aerodynamic resistance of the canopy, $r_{ac}$ is that of the canopy to in-canopy flow, $r_{aa}$ and $r_{as}$ are those of the reference height to in-canopy heat exchange plane height and there to the soil surface, respectively. $e_x$, $e_l$, $e_o$ and $e_s$ are the vapor pressures at the reference height, the leaf surface, the mean flow height, and the soil surface, respectively. $e_w (Ts)$ is the saturated vapor pressure at temperature $T_s$. 
$RI \ (= PR + TW - ET)$

Figure 5.6. Schematic diagram of runoff and infiltration processes in Sim-CYCLE. $RI$ is the inflow derived from precipitation $PR$, thawing snow $TW$ and evapotranspiration $ET$. $PN$ and $RO$ are the penetration and runoff, respectively. $MS$ is the soil moisture in the tank. $MS_{max}$ is the storage capacity in the tank. $\Delta MS$ is the increment of water storage in the unit time. $up$ and $lw$ indicate the upper and lower tank, respectively.
Figure 5.7. Seasonal changes of input climate data in the steady state experiment in (a) ground surface temp. \(T_{\text{surface}}\), 2m air temp. \(T_{\text{air}}\) and 10, and 200 cm soil temperature \(T_{\text{soil} 10}, T_{\text{soil} 200}\); (b) solar radiation at the top of atmosphere \(R_s\) top of atmosphere) and the ground surface \(R_s\) ground surface), cloudiness (C); (c) precipitation and specific humidity (q); and (d) soil water content at the depth of 10 cm \(T_{{\text{soil} 10}}\) and 200cm \(T_{{\text{soil} 200}}\), wind speed.
Figure 5.8. Growth of ecosystem carbon fluxes and storages, estimated by Sim-CYCLE equilibrium run. Results for the first 200 years. 1003 yrs are necessary for equilibrium.
Figure 5.9. Schematic diagram of carbon dynamics in the Haibei alpine meadow ecosystem, estimated by Sim-CYCLE equilibrium run.
Figure 5.10. Seasonal patterns of carbon fluxes, estimated by Sim-CYCLE equilibrium run
Figure 5.11. Seasonal patterns of carbon storages, estimated by Sim-CYCLE equilibrium run
Figure 5.12. Seasonal patterns of soil thermal profiles, estimated by Sim-CYCLE equilibrium run. Solid lines show the estimated profiles and broken lines show the measured profiles.
Figure 5.13. Seasonal patterns of heat budgets, estimated by Sim-CYCLE equilibrium run
Figure 5.14. Seasonal patterns of water budgets, estimated by Sim-CYCLE equilibrium run
Figure 5.15. Comparisons of the simulated above ground NPP (ANPP) of Sim-CYCLE transient run and observed values of other studies in the Haibei alpine meadow. RMSE = 0.089 Mg C m$^{-2}$ yr$^{-1}$. 
Figure 5. 16. Yearly changes of ecosystem carbon fluxes and storages, estimated by Sim-CYCLE transient run.
Figure 5.17. Annual climate and anomalies in carbon fluxes.
Figure 5.18. Relationships between temperature and carbon fluxes anomalies.
Figure 5.19. Relationships between precipitation and carbon fluxes anomalies.
Figure 5.20. Relationships between radiation and carbon fluxes anomalies.
Figure 5.21. Anomalies in monthly climate and carbon flux.
Figure 5.22. Seasonal changes in the slope of regression between temperature, precipitation, short wave radiation and carbon flux anomalies.
Figure 5.23. Transitional changes of carbon fluxes.
Figure 5.24. Transitional changes of carbon storages.
Figure 5.25. Relationship between prescribed environmental changes and equilibrium carbon fluxes
Figure 5.26. Relationship between prescribed environmental changes and equilibrium water, heat fluxes and soil water content.
Chapter 6 General Discussion

6.1. Carbon dioxide assimilation capacity of alpine meadow ecosystem on the Qinghai-Tibetan Plateau

The daily changes of CO2 exchanges showed the CO2 uptake in the daytime and CO2 release in the nighttime in summer and small CO2 release during all the days in winter in an alpine ecosystem (Fig. 3.30). The seasonal changes of CO2 exchanges showed the growing season that the ecosystem absorbed CO2 from May to September (Fig. 4.3). In 2002, the CO2 uptake during those five months was amounted to 138.4 g C m\(^{-2}\). The CO2 release during the other seven months was amounted to 59.9 g C m\(^{-2}\). The measurement provided very important information to estimate the annual carbon dynamics.

The maximum daily CO2 uptake (3.9 g C m\(^{-2}\) day\(^{-1}\)) was smaller than other grassland ecosystem at a similar latitude, although the maximum LAI was larger than those reported for these ecosystems (Sec. 4.4.3). The low temperature seemed limit the photosynthetic activity and thus the net CO2 uptake of the alpine ecosystem. The annual CO2 uptake (78.5 g C m\(^{-2}\) yr\(^{-1}\)) in 2002 was lower than those reported from other warmer ecosystems, e.g. temperate and tropical ecosystems, but was similar to those for other cool ecosystem, e.g. alpine and boreal forest ecosystems (Sec. 4.4.3). The low temperature shortened the growing season and thus may decrease the net CO2 uptake of the alpine meadow ecosystem.

The eddy covariance measurement indicated that this ecosystem was a CO2 sink, amounted to 78.5 g C m\(^{-2}\) yr\(^{-1}\) in 2002. On the other hand, the transition experiment for 1981-2000 using Sim-CYCLE showed that the annual NEP ranged from -70 to +70 g C m\(^{-2}\) yr\(^{-1}\) in the 20 years widely (Fig. 5.16). The measured annual CO2 uptake is close to the fluctuating range derived from model analysis, although this may contain the estimation errors caused by the energy imbalance and the gap-filling methods. However it is not enough to judge whether this ecosystem is a CO2 sink or source in average from the current measurement and model analysis. This is mainly due to the limited observation data. Furthermore multi-year data acquisitions of the CO2 exchanges are therefore required.
6.2. The response of the CO2 exchanges in alpine meadow ecosystem to the characteristic environments on the Qinghai-Tibetan Plateau

The Qinghai-Tibetan Plateau is characterized by extremely intensive sunlight, low temperature, little precipitation and limited plant growth. How do they affect the carbon dioxide uptake in this ecosystem? Which do they have a good influence or bad influence on it? In this chapter, the site-specific responses of CO2 exchange to four environmental factors, light, temperature, water and biology, were summarized and discussed in the followings.

**Light:** In the diurnal changes, the light responses of the afternoon CO2 uptake are lower than those of the beforenoon (Fig. 3.31 and 3.33). In the grassland ecosystems with low LAI, because of the low interception of intensive sunlight by vegetation, the soil surface temperature could be very high, which will further result in the increase of soil CO2 efflux as estimated in Fig. 3.35. In addition, the high VPD might cause the closure of stomata and thus suppress CO2 uptake. Such stomatal response to increased VPD has been observed in grassland (Verma et al., 1992) and forest ecosystems (Jarvis et al., 1997; Baldocchi and Vogel, 1997). In the alpine meadow, the increase of R_{canopy} is correlated with the decrease of ΔE but with the increase of VPD (Fig. 3.32). The amplitude of the diurnal SWC change was small and may play a very limited role in ecosystem respiration (Fig. 3.32b). The photosynthesis (i.e. GPP) in response to high PPFD did not differ between the beforenoon and the afternoon in this study (Fig. 3.35). It was concluded that the increasing air and soil temperatures accelerate plant respiration and soil organic matter decomposition, and as a result suppress the net CO2 uptake flux in this study.

Extremely intensive sunlight may induce down-regulation of photosynthesis in plants living in this alpine ecosystem. The depression of photosynthesis (i.e. GPP) increment in response to PPFD increasing was founded both in the beforenoon and the afternoon in this study (Fig. 3.35). Cui et al. (2003) investigated the leaf gas exchange and fluorescence emission in response to the changes in PPFD and leaf surface temperature for two Asteraceae herbaceous
species at same alpine meadow. They showed that under natural environmental conditions, apparent photoinhibition, indicated by reduced electron transport (ETR), was evident at high PPFD for both species. They also showed that the thermal dissipation, indicated by enhanced non-photochemical quenching (NPQ), which is associated with the xanthophylls cycle (Deming-Adams et al., 1996), played an essential role in photoprotection at low leaf temperature in the morning. At higher leaf temperatures, photorespiration was reported to play a more prominent role (Park et al., 1996; Streb et al., 1998; Manuel et al., 1999). Cui et al. (2003) showed the significant reduction of ETR in S. katochaete in the afternoon with high leaf temperature and expected that more than two-thirds of the photochemical energy was flown through the photorespiration pathway at a PPFD above 1800 µmol m⁻² s⁻¹ if there were no other significant electron acceptors. Thus, the photoinhibition and photoprotection was evident at the high PPFD for two Asteraceae herbaceous species, which is non-dominant species. Further research about the photoinhibition and photoprotection in dominant species is necessary to clarify the representative response to intensive sunlight.

In the seasonal changes, the daily NEP was correlated to daily PPFD positively in the growing season (Fig. 4.4), however, those linear relationships were slight weak. Gu et al. (in press, 2003) showed that for the same PPFD, the NEP was significantly higher on cloudy days than on clear days at the same site. Law et al. (2002) also founded that the NEP is more positive for cloudy condition than for clear sky condition in boreal aspen and Scots pine forests. Gu et al. (2002) suggested that diffuse radiation result in higher light use efficiency in plant canopies and has much less tendency to cause canopy photosynthesis. It is possible that the light quality caused the weakness of the linear relationship between daily NEP and PPFD. These findings call for different treatments of diffuse and direct radiation in models of ecosystem carbon cycle.

**Temperature:**

Respiration by autotrophs and heterotrophs was correlated positively with temperature in the seasonal changes (Fig. 4.5) and interannual changes (Fig. 5.18). The exponential coefficient
for the relation corresponds to a $Q_{10}$ of 3.21 (Fig. 4.5), higher than the value of 2.0 that is typically used as a default in modeling respiration (Law et al., 2002).

The annual GPP of 575 g C m$^{-2}$ (Table 4.5) was lower than those of boreal coniferous forest (723–959 g C m$^{-2}$ y$^{-1}$) and Colorado subalpine coniferous forest (831 g C m$^{-2}$ y$^{-1}$; 3050 m) at similar elevations, and much lower than that of tropical forest (3249 g C m$^{-2}$ y$^{-1}$), but within the range for temperate ecosystems, including forests and grasslands (542–1924 g C m$^{-2}$ y$^{-1}$; average, 1262 g C m$^{-2}$ y$^{-1}$; Falge et al., 2002). The daily GPP of the study site (3.59 g C m$^{-2}$ d$^{-1}$, Table 4.5) was similar to those of boreal evergreen forest and Colorado subalpine coniferous forest (4.6, 4.4 g C m$^{-2}$ d$^{-1}$, respectively), although slightly lower than those of temperate coniferous forest and C$_3$ crops and grassland (5.7–6.9 g C m$^{-2}$ d$^{-1}$). Thus, although our alpine meadow ecosystem has a daily CO$_2$ assimilation equal to that of a Colorado subalpine forest ecosystem, it has a lower annual GPP because of the restricted growing period, which was caused by low temperature.

The annual sum of NEP (78.5 g C m$^{-2}$ y$^{-1}$, Table 4.5) was close to that of the Colorado subalpine coniferous forest (71 g C m$^{-2}$ y$^{-1}$), although substantially lower than that of grassland (231.3 g C m$^{-2}$ y$^{-1}$) and boreal ecosystems (121.4 g C m$^{-2}$ y$^{-1}$). The daily NEP of the study site (0.49 g C m$^{-2}$ d$^{-1}$, Table 4.5) was similar to that of the Colorado subalpine forest (0.38 g C m$^{-2}$ d$^{-1}$). Although our alpine meadow ecosystem has a lower annual GPP than that of the subalpine forest ecosystem, it has a comparable annual NEP. It is assumed that not only low temperature but also small biomass suppresses the ecosystem respiration; as a result, this ecosystem may sequester substantial amount of C. Thus, much of the dynamic response of processes to temperature is lost in annual estimates, because of factors such as phonological influences.

**Moisture:**

In the seasonal changes, the CO$_2$ efflux increased as SWC decreased, and a CO$_2$ release of 0.94 µmol m$^{-2}$ s$^{-1}$ was maintained even in saturated conditions. This response to SWC has been observed in tundra under near-saturation (Oechel et al., 1998). Figure 4.6 might partly show the effect of increasing temperature. However, it is considered that the temperature
increment is not enough to increase respiration, and the depression of SWC from saturation to adequate contributes significantly to CO₂ efflux. This is an opposite response to that in relatively dry grassland (Hunt et al., 2002). It is assumed that the transition from aerobic condition (adequate moisture) to anaerobic condition (saturation) contributes to the rise in plant root and microbial activity. It suggests that ecosystem respiration in this meadow has an optimal SWC of approximately 0.3. Many respiration models mostly neglect the decline in microbial activity at high soil water content (Paul, 2001). So this result suggests that respiration depends on SWC.

In the diurnal changes and interannual changes, the apparent relationships between moisture factor and carbon dynamics were not founded. This suggested that alpine meadow ecosystem was advantaged in available water, although annual precipitation is not plentiful, because of low temperature.

Biology:

The linear regression slopes of NEP_{daytime} against PPFD changed with the changes in the LAI (Fig. 4.4). GPP_{max} is positively related to LAI (Fig. 4.10a). RUE_{GPP} and WUE_{GPP} also tended to increase as LAI increased (Figs. 4.10b, c). These results suggest that LAI determines the ecosystem capacity for assimilation and resource requirements in the seasonal changes. The LAI also seems the major factor causing the increase of ecosystem CO₂ uptake in interannual changes. The linear regression slopes of NEP_{daytime} against PPFD were higher for August 2001 than August 2002 (Fig. 4.4), probably due to higher CO₂ uptake rate and/or lower rate of ecosystem respiration. The LAI was 3.1 in August 2001 and 2.6 in August 2002.

In general, seasonal changes in respiratory processes are controlled by climate more strongly than by biological factors (Falge et al., 2002). However, at this study site, R_{c,max} which reached the maximum in autumn (Fig. 4.8), may be associated with above- and below-ground biomass, which reached the maximum at the same time (Fig. 3.24).

The maximum CO₂ uptake (−10.8 μmol m⁻² s⁻¹) at our site 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 uptake potential with lower release potential. The Kobresia 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 due to the low shoot system in the alpine meadow, which resulted from the relatively greater abundance of broad-leaved species and the low canopy height. The high LAI, on the other hand, suggests a likely contribution to the high belowground biomass in the alpine meadow. The belowground biomass was 6-7 times higher than the aboveground biomass (e.g. 1892 g d.w. m\(^{-2}\) on 11 August 2001; unpublished data). This particular allocation pattern of plants may favor high soil carbon storage in the alpine ecosystem. Although the CO\(_2\) uptake was not high, the daily net ecosystem carbon gain 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 ecosystem respiration. It also could be the low maintenance respiration due to low aboveground biomass in the alpine meadow (Table 3.2).

The maximum \(\text{NEP}_{\text{total}}\) of our site (3.9 g C m\(^{-2}\) d\(^{-1}\); Table 3.2) was 20-55% less than those of tallgrass prairies in Kansas and Oklahoma, USA (4.9-8.4 g C m\(^{-2}\) d\(^{-1}\), Kim et al., 1992; Ham and Knapp, 1998; Suyker and Verma, 2001). However, seasonal maximum was almost four times larger than that of a subalpine conifer forest in Colorado (1.0 g C m\(^{-2}\) d\(^{-1}\); Monson et al., 2002) at similar altitude (~3050 m). As pointed out earlier, not only low temperature but also small biomass significantly suppressed plant respiration at our site.

6.3. The global warming effects on the ecosystem carbon sequestration

The model sensitivity analysis showed that the \(GPP\), \(AR\), \(NPP\) and \(W_{\text{plant}}\) responded quickly, while the \(HR\) and \(W_{\text{soil}}\) responded very slowly against the climate change. These may indicate that the long-term global warming will increase the ecosystem carbon uptake. Cao and Woodward (1998) used a terrestrial biogeochemical model, forced by simulations of transient
climate change with a general circulation model to quantify the dynamics variations in ecosystem carbon fluxes induced by transient changes in atmospheric CO2 and climate from 1861 to 2070. They predicted that these changes increased global net ecosystem production significantly, but that will decline as the CO2 fertilization effect becomes saturated and is diminished by changes in climate factors. Cox et al. (2000) presented results from a fully coupled, three-dimensional carbon-cycle model, indicating that carbon-cycle feedbacks could significantly accelerate climate change over twenty-first century. They found that under a business as usual scenario, the terrestrial biosphere acted as an over all carbon sink until about 2050, but turn into a source thereafter. Thus, both in alpine meadow and in global scales, the long-term global warming seems to make the terrestrial ecosystems as a carbon sink at least in its initial period (i.e. ~ 2050), but that seems decline as the CO2 fertilization effect is saturated.

The increments of temperature by 5 °C increased the GPP but the increase over 7.5 °C decreased the GPP. Zhang and Welker (1996) showed that while the peak community biomass showed no significant change, the duration of peak biomass was extended in the warming experiment by 5 °C of air temperature at the Haibei alpine meadow. Thus, the warming by 5 °C advances the ecosystem photosynthetic activity and increases the duration of growing season, and thus, the GPP. However, the temperature increase over 7.5 °C may go beyond the optimum temperature of photosynthesis for the objective ecosystem and decreases the GPP. This suggests that the ecosystem may shift to another biome type that would survive the high temperature environment. In this analysis, the optimum temperature of photosynthesis plays an important role to determine the threshold temperature to change the warming effects on the ecosystem canopy growth from positive to negative.

To obtain the ecosystem response to the climate change qualitatively and quantitatively, shorter- and longer-term measurement about the ecosystem carbon dynamics are necessary.

6.4. Recommendations for future research

1) The validation of hypothesis in the several environmental controls on the CO2 exchanges.
To clarify the mechanism involved in the environmental controls on CO2 dynamics in the alpine ecosystem, for examples, the CO2 uptake depression in the afternoon at the same light intensity, the smaller scale measurements on single leaf are necessary.

2) The long-term eddy covariance measurement of the CO2 exchanges

To assess the average and fluctuation of the CO2 exchange in this ecosystem, the CO2 flux measurement using the eddy covariance method is necessary to continue for several years.

3) The model construction using the measured ecophysiological parameters

To construct the representative model to simulate the ecosystem carbon dynamics, the ecophysiological parameters are necessary to be measured at the experimental sites. In this study, Sim-CYCLE used the ecophysiological parameter provided for global scale simulation.

4) The longer-term model simulation to investigate the effect of global warming

In general, there are three experimental styles in the model simulation studies; the spin-up experiment, the historical experiment and the future experiment.

The spin-up experiment is conducted to equilibrate the carbon dynamics using the climate data before 1850s, in which the ecosystem reached the climatic climax. The historical experiment is conducted to investigate the past climate perturbation effect, containing the global warming, and the ecosystem carbon dynamics using the re-analyzed climate data. The future experiment is conducted to predict the climate change effect on the ecosystem carbon dynamics using the future modeled climate data induced by the global warming scenarios.

In this ecosystem, these three experiments are necessary to represent the more real ecosystem response to climate changes in the long-term scale.
Chapter 7 Conclusions

Considering the vast area of about $2.5 \times 10^6$ 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. The alpine meadow is located at either the horizontal extreme of the plateau, or at a low altitude within the plateau with averaged elevation of above 4000m. However, the alpine Kobresia meadow ecosystem is one of the most widely distributed vegetation on the Qinghai-Tibetan Plateau (Zhou, 2001). The alpine meadow occurs in the areas with a large elevation ranging from 3200 to 5200 m with variable climatic conditions. If it can be assumed that carbon budget depend on more on vegetation types than on any particular geological conditions, it is thus conclusive that the knowledge obtained from the current ecosystem should provide an important insight into our understanding on the carbon dynamics for the grassland ecosystems on the Qinghai-Tibetan Plateau. It should be noticed here that various grassland ecosystems occupy more than 60% of the plateau (Wang et al., 2002).

The CO2 flux measurement was conducted for two years by the eddy covariance method in an alpine Kobresia meadow on the Qinghai-Tibetan Plateau. Such the measurement provided the first example of CO2 exchange in the blank area of global flux network in an extreme environment from China. The evidence from the observation led to the conclusion that the alpine meadow was a CO2 sink at least in 2002. The current CO2 sink strength seems not too high (78.5 g C m$^{-2}$ yr$^{-1}$), but is comparable with many sub-alpine ecosystems reported so far. Moreover, it is also concluded that temperature was the major environmental control on CO2 exchange and the low temperature limited evidently the ecosystem respiration. Soil water availability was high in the alpine meadow. The high soil water might reduce the ecosystem respiration. This conclusion seems subjected to the argument that it is difficult to separate the effects of the temperature and soil water contents on the ecosystem carbon budget. However, it is evident that the soil water content showed very sharp temporal variations during a short period such as one week or so, while temperature exhibited much gentle and small variation at the similar temporal scale. Therefore, it is concluded that a high soil water content is correlated
with a low ecosystem respiration at nighttime for the alpine ecosystem.

The CO₂ flux was modeled by the ecosystem carbon dynamics model, Sim-CYCLE, for 20 years at the same site. The modeling provided the reasonable interannual changes of CO₂ exchange. The results of modeling led to conclude that annual carbon budgets were fluctuated ranging from +70 to -70 g C m⁻² yr⁻¹ and it was unclear whether this ecosystem was a carbon sink or source for long-term periods qualitatively. It is also concluded that temperature was the major environmental control on annual CO₂ exchange similarly to the measurement results. The model sensitivity analysis suggested that the long-term global warming by 5 °C in annual average temperature would increase the ecosystem carbon uptake due to the extension of the ecosystem photosynthetic activity and the duration of growing season. However, the long-term global warming over 7.5 °C may go beyond the optimum temperature of photosynthesis for the objective ecosystem. This suggests that the ecosystem may shift to another biome type that would survive the high temperature environment.

The alpine meadow exhibited a relatively low ecosystem CO₂ uptake when taken into account of the fairly high leaf area index of about 3. From this study, it is further concluded that the alpine meadow has the potential to sequester carbon, but the potential appears to be small possibly because of the limitation of low temperature. Further studies are needed to clarify the environmental controls on the carbon dynamics of the alpine meadow ecosystem from in short-term to in long-term periods.
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