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Agrometeorological conditions of grassland vegetation in central Mongolia and their impact for leaf area growth

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[1] The long-term observation of surface heat and water budget and hydrometeorological elements has been carried out over a grassland site at Arvaikheer (46.23°N, 102.82°E) in central Mongolia as the framework of the GEWEX Asian Monsoon Experiment-Asian Automatic Weather Station Network (GAME-AAN). The purpose of this study is to clarify the relationship between vegetation and climate using long-term data (1982–2000) of satellite-derived leaf area index (LAI) and climatic data observed at Arvaikheer. Furthermore, we aimed to reveal physical process by comparing soil moisture and heat and water budgets in 1999 and 2000 as a case study of good and poor vegetation growth. Significant positive correlations with 99% confidence levels were found for July precipitation (P) and the LAI in July (0.538), August (0.826), and September (0.564). Composite analysis for five highest (H5) and lowest (L5) LAI years showed the significant positive anomalies of P in July and LAI in July and August for H5. In June and July 1999, soil moisture and P values were higher than values in 2000; this pattern was reversed in August and September. The mean LAI during the 1999 growing season (1.0) was about twice that of 2000 (0.6). In 1999 the ratio of evapotranspiration (ET) to P (ET/P) and change of stored soil moisture (ΔW) to P (ΔW/P) were 0.79 and 0.15, respectively. In 2000, ET/P and ΔW/P were 0.94 and 0.0, respectively. These results suggest that the P and ΔW before July had the most influent on grass growth in central Mongolia. *INDEX TERMS:* 1818 Hydrology: Evapotranspiration; 1866 Hydrology: Soil moisture; 3309 Meteorology and Atmospheric Dynamics: Climatology (1620); 3322 Meteorology and Atmospheric Dynamics: Land/atmosphere interactions; 3360 Meteorology and Atmospheric Dynamics: Remote sensing; *KEYWORDS:* Mongolia, grassland, leaf area index, soil moisture, rainfall, evapotranspiration

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1. Introduction

[2] The global average surface temperature has increased over the 20th century by about 0.6°C [*Intergovernmental Panel on Climate Change* (*IPCC*), 2001]. Most warming has been observed over midlatitude and high-latitude areas of Asia and parts of western Canada [*IPCC*, 2001]. Over a recent 40-year period (1951–1990), northern regions of China and Mongolia have shown remarkable linear increases in annual mean temperature [*Yatagai and Yasunari*, 1994]. Warmer winter and spring temperatures mainly account for these trends [*Yatagai and Yasunari*, 1994]. Applying rotated empirical orthogonal functions (REOF) to a recent 40-year period of summer precipitation data also showed a significant decreasing trend after 1955 from north China to central and southeastern Mongolia [*Yatagai and Yasunari*, 1995].

[3] Few studies have examined the relationship between vegetation and climate in Mongolia. *Kondoh and Kaihotsu* [2003] found that Mongolian grasslands showed a clear relationship between total summer precipitation and the NDVI amplitude. *Ni* [2003] noted that grasses showed a positive relationship to precipitation and aridity index (mean annual precipitation divided by mean air temperature plus 10 degrees) in northeast China and southeast Mongolia. *Suzuki et al.* [2003] pointed out that the Mongolia showed 11 weeks later start of green-up than Kazakh, while Mongolia and Kazakh located in almost same latitude. They suggested that Kazakh had more soil moisture from snow-

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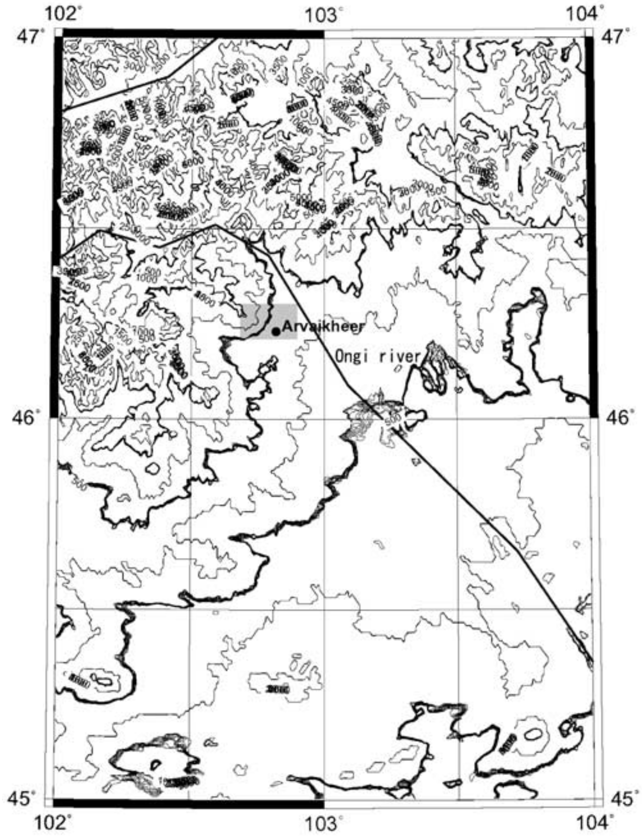

Figure 1. Study site location. Contour lines show the topography around the site. Closed circle shows the observational site. Shaded rectangle shows the area of the LAI data. The topographic data is derived from GTOPO30.

melt than its in Mongolia. However, neither study clarified physical or physiological processes together with intersummer variation of precipitation and vegetation.

[4] The long-term observation of surface heat and water budget in association with hydrometeorological elements has been carried out over a grassland site at Arvaikheer (46.23°N, 102.82°E) in central Mongolia as the framework of the GAME-AAN (GEWEX Asian Monsoon Experiment-Asian Automatic Weather Station Network; *Sugita et al.* [2003]). The purpose of this study is to clarify the relationship between vegetation and seasonal and interannual variability of rainfall using long-term data (1982–2000) of satellite-derived LAI and climate variables observed at Arvaikheer. Furthermore, we aimed to reveal physical process by comparing surface climate, soil moisture, and heat and water budgets of in 1999 and 2000 as a case study of good and poor vegetation growth.

2. Data and Methods

2.1. Observational Site Description and Climatic Data

[5] Long-term observations of hydrometeorological elements and surface flux have been conducted by using Automatic Weather Station (AWS) at the site in Arvaikheer in central Mongolia (closed circle shown in Figure 1) since 17 September 1997. The area features grassland vegetation in summer and snow cover in winter and is classified as dry steppe [*Gunin et al.*, 1999]. Arvaikheer is located in the alluvial fan of Ongi River which flows from Khangai mountains to the Gobi desert. Annual mean air temperature and annual precipitation (P) are 0.4°C and 245 mm, respectively. The study site is located at the southern edge of Arvaikheer Airport, which has an unpaved, grass-covered runway. The mobile observational result of soil moisture around site shows the spatial distribution of soil moisture was monotonous at the 5–10 cm depth in 10 km by 16 km around the site except the area along Ongin River [*Kaihotsu et al.*, 2000]. All the vegetation at this site is C_3-type grasses, mainly in the Gramineae (e.g., *Stipa Gobica*), Compositae (e.g., *Artemisia adamsii*), and Cyperaceae (e.g., *Carex duriuscula*) families. The soil consists of brown humus and sand from the surface to 20 cm, brown soil with rock from 20–70 cm, and brown silt below 70 cm. Winds prevail from the northwest and homogeneous grasslands extend more than 5 km around the site. Surface flux was observed at 7.8 m height by Portable Automated Mesonet III (PAM III; *Militzer et al.* [1995]) Automatic Weather Station (AWS) developed by the National Center for Atmospheric Research (NCAR) in the United States. The observational height of surface flux to ensure sufficient fetch since surface fluxes are considered 100 times greater than the measuring height (780 m; *Horst and Weil* [1994]). The details about the observational system are described in Appendix A.

[6] Mongolia's Institute of Meteorology and Hydrology (IMH) provided daily P data from 1999 to 2000 at Arvaikheer. The IMH also provided long-term 10-day mean air temperature and 10-day P data from 1982 to 2000. In addition to station data provided by IMH, we used summarized global data on daily mean air temperature and daily precipitation amounts. These data were obtained from the Global Climate Observing System (GCOS) Surface Network (GSN), produced by the National Climate Data Center (NCDC), and are available via the NCDC web server (http://www.ncdc.noaa.gov/). The topographic data (GTOPO30; Global 30 arc second elevation data) used for Figure 1 was distributed by the Land Processes Distributed Active Archive Center (LP DAAC), located at the U.S. Geological Survey's EROS Data Center http://LPDAAC.usgs.gov.

2.2. Remote Sensing Data

[7] In this study, we used a global LAI data set derived from global composites of maximum NDVI values. The data are part of the National Oceanic and Atmospheric Administration (NOAA) Pathfinder Advanced Very High Resolution Radiometer (AVHRR) Land data set (PAL) version 3. The global composites were created using an algorithm that incorporates results from a three-dimensional radiative transfer model and a six-biome classification scheme, as described in the work of *Myneni et al.* [1997]. The data contain monthly LAI values at a 16 × 16 km spatial resolution from July 1981 to May 2001 and are available online at http://cybele.bu.edu.

[8] The NDVI data set used for deriving the LAI was corrected in three stages, as further described in the work of *Nemani et al.* [2003]. *Buermann et al.* [2002] examined four kinds of LAI validation based on an algorithm *Myneni et al.*

[1997] derived from NDVI values. The LAI data set used here is the same as that used in the work of *Buermann et al.* [2002]. In their study, relative error in the LAI was about 10–20% in dense biomes, which is approximately equal to the estimated mean uncertainty in the AVHRR channel data from which the NDVI is computed. Satellite sensor-derived LAI values were comparable to field observations at representative grass and needle forest biome sites in *Buermann et al.*'s [2002] analysis. Their LAI data showed good qualitative agreement with one other LAI data set, but was consistently larger than another LAI data set. Interannual variation in their data set was meaningful, because of the spatial and temporal agreement between the satellite LAI and station rainfall anomalies in tropical regions affected by the ENSO phenomenon. Consequently, we considered the global data used in this study to be sufficient for discussing interannual LAI variability. We used the data from the grid near Arvaikheer (46.175–46.325°N, 102.66–102.86°E; shaded area in Figure 1).

3. Relationships Between Local Climate and LAI

[9] According to analysis of long-term monthly climatic statistics using simple bioclimatic indices, water availability most strongly limits vegetation growth over 40% of Earth's vegetated surface; temperature and radiation most limit growth over another 33% and 27% of Earth's vegetated surface, respectively [*Nemani et al.*, 2003]. According to *Nemani et al.* [2003], Mongolia is located near the boundary of the temperature-limited and water-limited regions. The monthly mean air temperature (T) at Arvaikheer exceeds 5°C in May and the monthly mean T drops lower than 5°C in October. It is common that the monthly mean T with 5 degrees is the threshold for vegetation activity. Moreover, most precipitation (about 90% of annual precipitation) and growth of vegetation at Arvaikheer occurs from May to September. Therefore we define the vegetation growing season (GS) as May to September.

[10] To clarify the seasonal variation of T, P and LAI at the site, we present the time series of monthly T, P and LAI averaged from 1982 to 2000 with standard deviations (SD) in Figure 2. The seasonal maximum of both T (16.1°C) and P (77.0 mm) appeared in July, while the seasonal minimum of both T (8.9°C) and P (14.7 mm) appeared in September. P had large SD from June to August with more than 30 mm, while T had small SD from June to August with less than 1.5 degrees. LAI increased from May (0.4) to June (1.2) rapidly then became nearly constant until September with slight maximum in August (1.3). From 1982 to 2000, T during GS had significant increasing trend (0.1 degree/year) with 99% confidence level. In same period, P and LAI during GS had no significant trend.

[11] To evaluate the relation between climate and vegetation, it is useful to calculate correlation between T, P and LAI. No significant correlation showed for T and LAI. Table 1 shows statistical correlations between P and the LAI from 1982 to 2000. Nineteen data points were used for this analysis. Thus the degree of freedom for correlation analysis was 17, and the lower limits for significant correlation were 0.529 and 0.456 at 99% and 95% confidence levels, respectively. Positive significant correlations at 99% confidence levels were found for P in July and the LAI in July (0.538), August (0.826) and September (0.564). Positive significant correlations at 95% confidence levels were found for P in May and the LAI in June, and for P in June and the LAI in June. No correlation showed for P in August and the LAI in August or September, or for P in September and the LAI in September, implying that after August, P has less effect on grass growth.

[12] To reveal the actual phenomena guiding relationships between P and the LAI, we created the composites of years with the five highest (H5; in the years of 1984, 1985, 1990, 1997, and 1998) and five lowest (L5; in the years of 1982, 1986, 1988, 1995, and 2000) for mean LAI value during GS. We present the anomalies of P and LAI for comparing H5 and L5 composite in Figure 3. P of H5 in July had significant (larger than SD) positive anomaly with 45.4 mm while P of L5 had nearly significant negative anomaly with 44.4 mm. The LAI of H5 in July and August had the significant positive anomaly with 0.7 and 0.6, while the LAI of L5 from July to September had the almost significant negative anomaly with 0.5. In other months, no significant difference showed for P and the LAI. These results support findings of large, significant positive correlations for P in July and the LAI from July to September. This suggests that plenty of July rainfall is the most important factor for grass growth.

4. Comparison of Summer Climate and Vegetation in 1999 and 2000

4.1. Hydrometeorological Condition

[13] Annual P in both years (150 mm) was about 60% of that in normal years (240 mm) while seasonal variability in P differed in these years. We used these two years in a case study to examine results from the previous section further. Figure 4 shows the time series of daily P from May to September in 1999 and 2000. From May to July, 1999 had a higher P frequency (43.5% of total days) than 2000 (33% of total days). In 1999, P values exceeding 3 mm/day occurred on 13 days from May to July; only 7 days in the same period in 2000 exceeded 3 mm/day. From August to September, both 1999 and 2000 showed the same frequency of P. In 1999, no day had a P value greater than 10 mm/day; four such days were observed in 2000.

[14] The surface condition is the most important parameter controlling heat and water budgets. Soil moisture is a useful parameter for expressing the surface condition. To discuss mechanisms of evaporation from the soil and rainfall infiltration into the soil, information on soil physical parameters and a moisture characteristic curve are useful. Table 2 shows laboratory measurements of soil physical parameters at Arvaikheer. It suggests that the near surface soil had a greater water-holding capability than deeper soil. Figure 5 shows the moisture-characteristic curve of soil at Arvaikheer. Values were derived from laboratory experiments using the suction plate assembly method. At the same volumetric soil water content, soil from 10-cm depths showed lower pressure head values than deeper soil. This implies that the soil at 10 cm depth has stronger suction powers than the deeper soil layer. Owing to the limitation of laboratory experiment, we could not get the moisture characteristic curve of soil for the pressure head value higher than 450. For the soil at 10 cm and 55 cm, it is not enough to describe the behavior of the

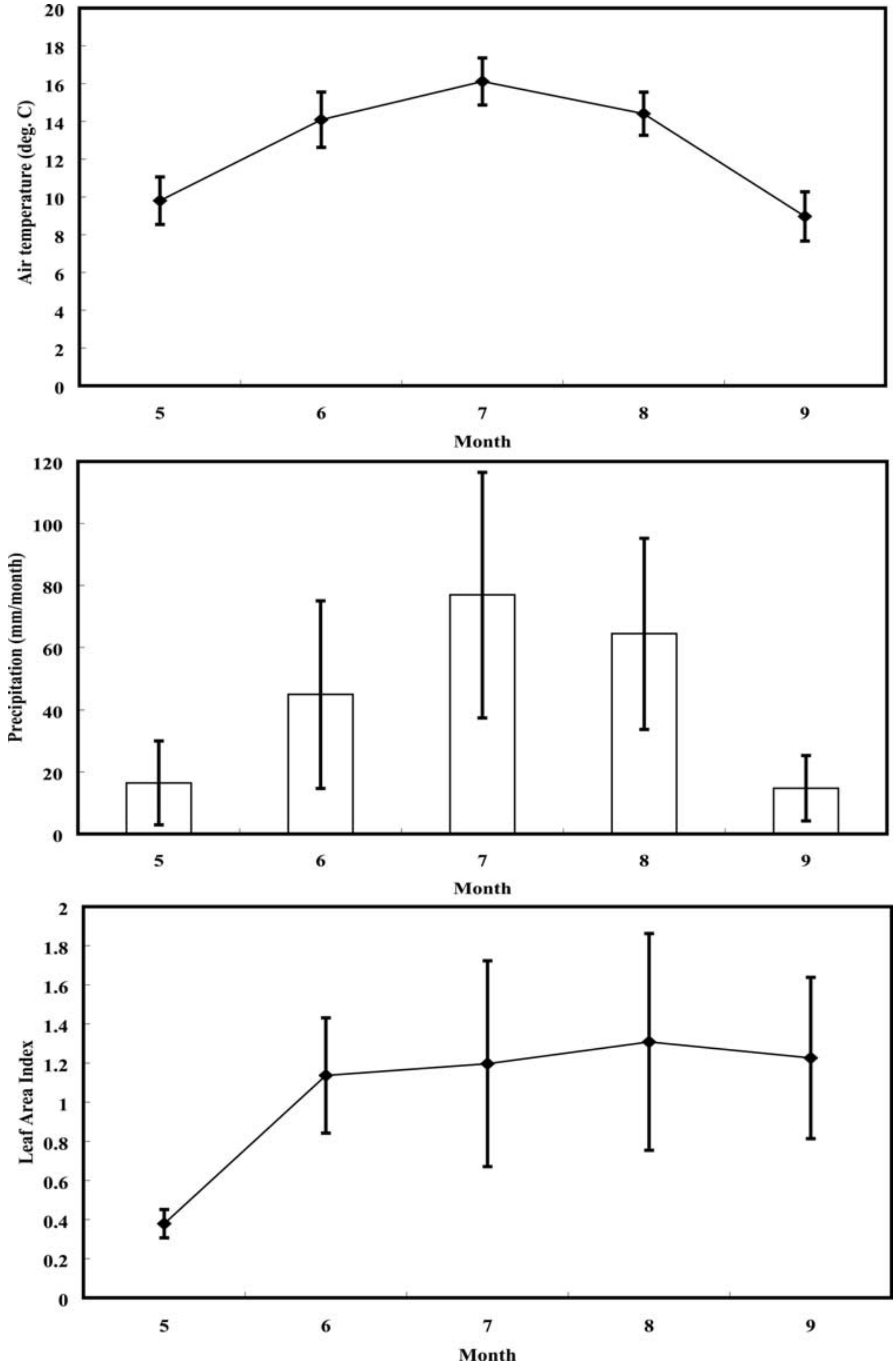

Figure 2. Time series of (top) monthly mean air temperatures, (middle) monthly accumulated precipitation, and (bottom) leaf area index averaged from 1982 to 2000. Error bars show the standard deviation from 1982 to 2000.

soil drier than 0.2 of the volumetric soil water content (θ). As the wilting point and filed capacity might be less than 0.2, we did not have the data of them for the soil at Arvaikheer. For the reference, *Erdenetsetseg* [1997] showed the filed capacity of soil in steppe region for upper soil layer to 50 cm depth was from 0.1 to 0.15.

[15] Figure 6 shows the time series of 5-day mean θ at depths from 5 to 20 cm, on average, during the growing season. In both 1999 and 2000, θ responded to P very well. The θ in 1999 was higher than in 2000 in June and July, and lower than in May and August. *Erdenetsetseg* [1997] showed the soil moisture regime at typical sites with long-

Table 1. Correlation Matrix of P and the LAI[a]

	LAI				
P Month	5	6	7	8	9
5	0.245	**0.482**[b]	0.074	0.069	0.086
6	—	**0.458**[b]	0.445	0.050	0.053
7	—	—	**0.538**[c]	**0.826**[c]	**0.564**[c]
8	—	—	—	−0.141	−0.065
9	—	—	—	—	−0.009

[a]The bold values are significant correlation coefficient.
[b]The 95% significance level.
[c]The 99% significance level.

term observation. The temporal variation of θ in 1999 resembled with the long-term mean soil water content in steppe region, while the temporal variation of θ in 2000 was similar to that in desert region shown in the work of *Erdenetsetseg* [1997]. As shown in Figure 4, the frequency of P in 1999 was about 1.5 times of its in 2000. This also affected the different intraseasonal variation of θ between in 1999 and 2000. These results are consistent with results presented in the work of *Knapp et al.* [2002], in which the mean soil moisture content was reduced when the rainfall interval was extended.

[16] To consider the transpiration from the leaves of grass, deeper soil moisture is important. As shown in the work of *Gunin et al.* [1999], most of the root of grass was located from several centimeters to one meter especially many roots was found below depths of 30 cm. Figure 7 shows the time series of 5-day mean volumetric soil water contents at depths from 5 to 95 cm during the 1999 and 2000 GSs. In mid-July of 1999, θ at depths of 40 cm, 70 cm and 95 cm abruptly increased from 0.09 to 0.15 (depth of 40 cm) followed the increase of θ at depths of 5 cm and 10 cm. In 2000, θ at depth of 40 cm, 70 cm and 95 cm did not show any change during GS. These results show that there was no water infiltration to deeper layer in 2000. It is likely that most of rainfall was evaporated as soon as rainfall occurred in 2000 and did not infiltrate to deeper soil layer, because the interval of P was large and amount of P was small. In contrast, much water infiltrated to deep soil layer below depths of 40 cm in 1999, which there was much water amount for the grass to suck the water from deeper soil through their roots.

[17] Figure 8 shows a time series of the monthly LAI derived from satellite-sensor NDVI observations over Arvaikheer. The mean LAI during the 1999 GS (1.0) was about twice that in 2000 (0.6). The annual maximum LAI value reached 1.6 in June 1999 and 0.7 in June 2000. The uncertainty in the LAI estimate may be order of 0.5 LAI [*Myneni et al.*, 1997]. Therefore the difference of annual maximum LAI is significant. The LAI peak in June 1999 likely related to the θ in root zone deeper than 10 cm in June. The θ at 10 cm and 20 cm in June of 1999 were about twice of its in June 2000 (Figure 7). In May and June 1999, the frequency of P was about twice of its in May and June 2000 (Figure 4). The days with P more than 5 mm/day were four times in middle and late May 1999 while there was no day with P more than 5 mm/day in May and June 2000 (Figure 4). In June and July 1999, θ values were higher than in 2000, while θ values in August and September 2000 were higher than in 1999. P in June and July 2000 were significant smaller than the average P from 1982 to 2000. It implies that the small P caused low θ in June and July 2000. These results point out that higher θ in early 1999 GS may have an impact on the LAI since the 1999 LAI was significantly greater than the 2000 LAI.

4.2. Surface Heat Budget

[18] The available energy (Net radiation (Rn) minus soil heat flux (G); Ra = Rn − G) is the most important for controlling the heat budget. Figure 9 shows the time series of 5-day mean Ra during the GS in 1999 and 2000. From May to early July in 1999, Ra was about 7 MJm^{-2}/day smaller than in 2000 except some period. From mid-July to September, Ra in 1999 and 2000 were almost same. The lower values in 1999 correspond to successive rainy days.

[19] To test flux measurement performances, we checked surface heat balance closure by comparing the sum of the turbulent fluxes (TF) with Ra. Regression slopes were 0.9326 and 0.9043 in 1999 and 2000, respectively, suggesting that TF was about 10% lower than Ra. Such a TF shortfall is often called an energy imbalance [*Panin et al.*, 1998]. Many studies have examined the cause of energy imbalances [e.g., *Gu et al.*, 1999; *Mahrt*, 1998; *Lee*, 1998; *Tsukamoto et al.*, 1995; *Twine et al.*, 2000; *Tanaka et al.*, 2001; *Toda et al.*, 2002; *Wilson et al.*, 2002]. Besides measurement error, most studies have noted that an energy imbalance results from the effects of nonzero mean vertical velocity, nonstationary measured time series, and the development of horizontal energy advections in association with horizontal inhomogeneity. The energy imbalance observed in this study was less than 10%, which is same as the probable error of Ra for homogeneous sites showed by *Twine et al.* [2000]. It is still difficult to evaluate the causes of energy imbalances that are less than 10% of Ra. The energy imbalance less than 10% is quite better than the mean values observed in 22 sites of FLUXNET was in order of 20% [*Wilson et al.*, 2002]. Thus the turbulent fluxes observed in this study may be considered accurate enough for quantitative analysis.

[20] To consider heat budget characteristics, we divided available energy into sensible heat flux (H) and latent heat flux (LE). Figure 10 shows a time series of 5-day mean sensible heat flux and latent heat flux during the GS. From May to June, H was the dominant component of the surface heat budget with about 15 to 20 MJm^{-2}/day in both years; during this same period, H reached its annual maximum of 20.2 MJm^{-2}/day and 22.3 MJm^{-2}/day in early June 1999 and mid-June 2000, respectively. In 1999, LE exceeded H in three periods (mid-June, early August, and late September), in response to soil moisture increases after heavy rainfall. In 2000, LE surpassed H in mid-June and early August during high soil moisture conditions. In 1999, LE reached its annual maximum in mid-July at about 14.7 MJm^{-2}/day that was about 60% of Rn, while LE in 2000 attained an annual maximum in early August at about 12.4 MJm^{-2}/day that was about 50% of Rn. This result is different from the results of Japanese grassland [*Saigusa et al.*, 1998] which the ratio of LE to Rn was same between wet and dry year. This difference may imply that the Mongolian grassland is more sensitive for the difference of climate than the Japanese grassland where there is enough precipitation for the grass growth.

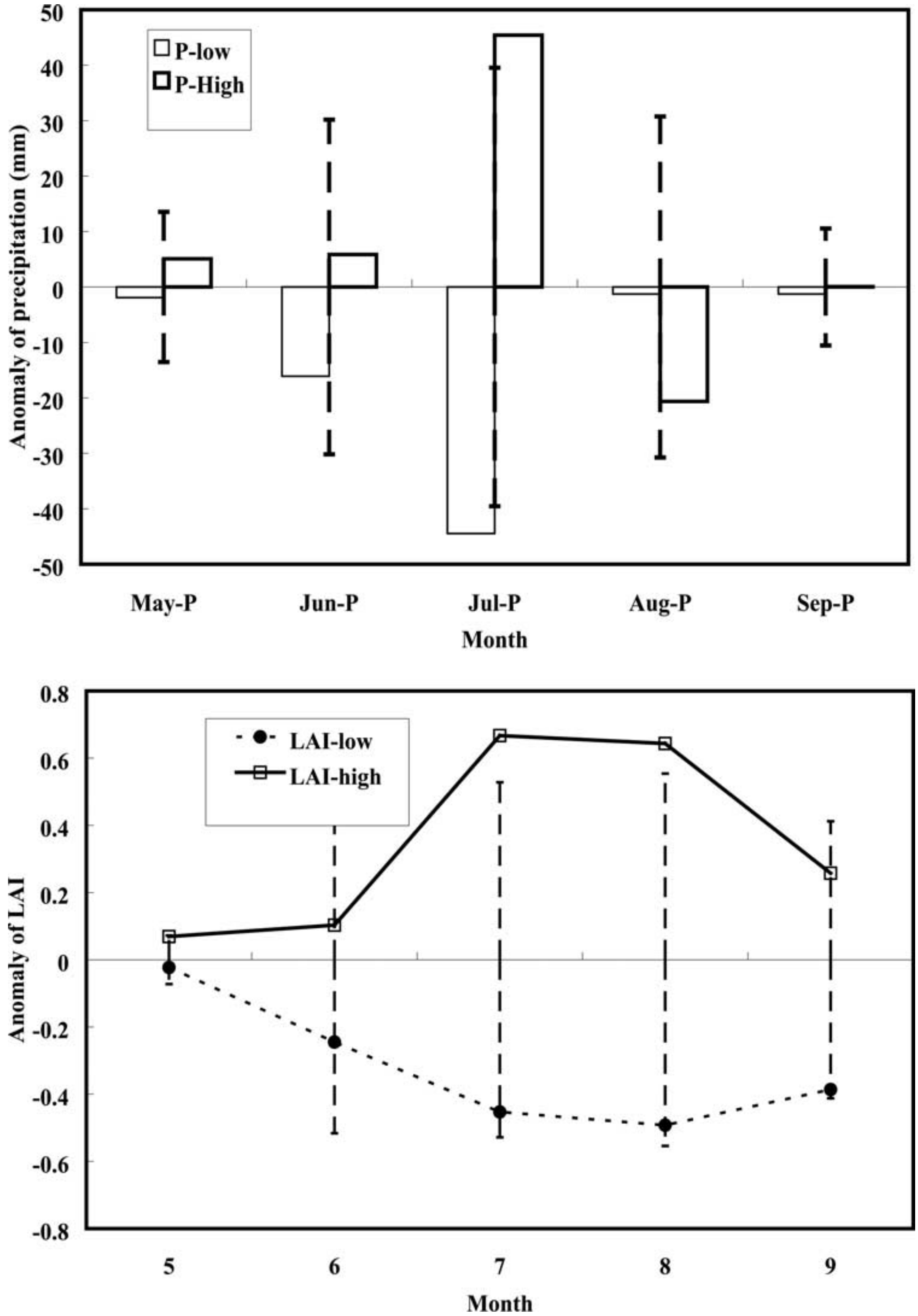

Figure 3. Comparison of (top) P (bottom) LAI by composites of the five years with the highest LAI values and 5 years with the lowest LAI values, which was chosen from mean LAI value during growing season. Error bars show the standard deviation over 19 years (1982–200).

[21] The equilibrium latent heat flux (LE_{eq}) is defined as

$$LE_{eq} = \frac{\Delta}{\Delta + \gamma}(Rn - G), \qquad (1)$$

where Δ and γ are slope of the saturation water vapor pressure curve and the psychrometric constant [*Brutsaert*, 1982]. The LE/LE_{eq} is often called as Priesley-Taylor coefficient. Typically, $LE/LE_{eq} < 1$ represents a surface where limitations in water supply are sufficient to reduce LE below the equilibrium evaporation rate; in contrast, $LE/LE_{eq} > 1$ typifies surface where the water supply is unrestricted and available energy limits evaporation [*Blanken et al.*, 1997]. Figure 11 shows the time series of the 5-day mean LE/LE_{eq} during the 1999 and 2000 growing seasons. In 1999, LE/LE_{eq} varied from 0.1 to 1.1; the average value during the GS was 0.38. Values ranged from 0.13 to 0.72, and the average value was 0.37 in 2000. The

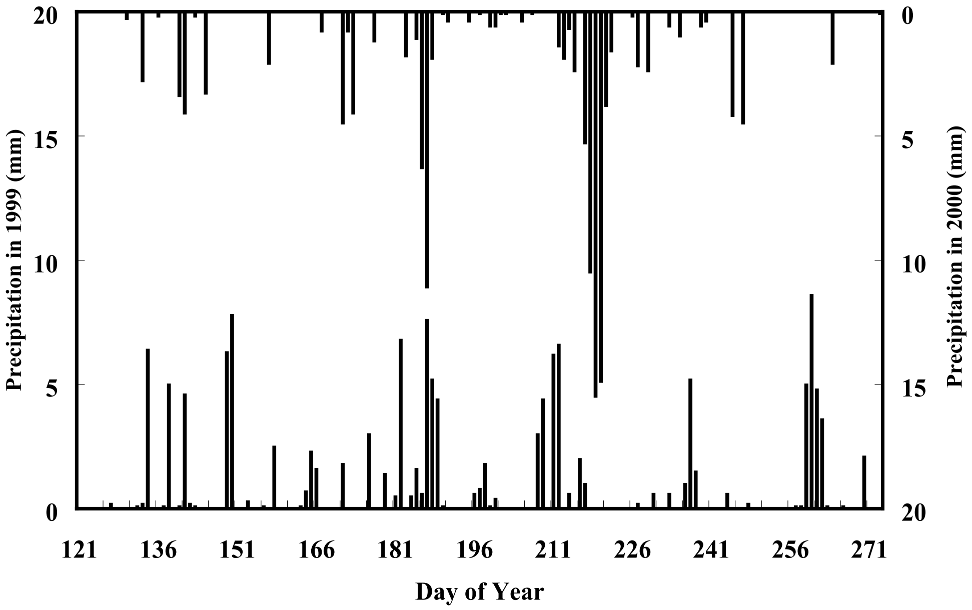

Figure 4. Comparison of daily precipitation from May to September in 1999 and 2000.

mean LE/LE_{eq} during the first half of the GS (May to mid-July) in 1999 (0.34) was higher than that in 2000 (0.28); the mean LE/LE_{eq} during the latter half of the GS (mid-July to September) in 1999 (0.42) was lower than that in 2000 (0.46). The difference of LE/LE_{eq} between years in 1999 and 2000 is well corresponded to the difference of θ (Figure 6) in both years. The 1999 peak (1.1) at our study site resembled results from a temperate grassland in northern Canada (0.95) in a wet year (1998; *Wever et al.* [2002]). However, the lowest LE/LE_{eq} value in our study (0.1) was much smaller than that in the Canadian study (0.3). It is likely that water availability exerts a stronger control at our site.

4.3. Surface Water Budget

[22] As described in the previous section, most of the data used in this analysis were daytime mean values from 0900 to 1600 LST. However, to examine the water budget, we needed to use mean daily values. We estimated daily evaporation using instantaneous or multiple observations of evaporative fractions (EF = LE/(Rn − G)), under an assumption proposed by *Sugita and Brutsaert* [1991]. *Sugita and Brutsaert* [1991] determined daytime evaporation as

$$LE_d = EF_d Q_d, \tag{2}$$

where LE_d is daytime evaporation, EF_d is the daytime evaporative fraction, and Q_d is the daytime net radiation. They argued that this equation would not apply to the estimation of daily evaporation since the assumption of constant EF is not satisfied at nighttime. Thus for practical purposes they proposed a simple correction procedure:

$$ET_{td} = aET_d, \tag{3}$$

where ET_{td} is the total daily evapotranspiration, ET_d is the daytime evapotranspiration, and "a" is a locally calibrated constant.

[23] In our study, daily evaporation is obtained by a = 1.05. The correlation coefficient of between ET_{td} and ET_d is quite high (r = 0.996); the slope through the origin and the standard error of estimation are 0.9463 and 0.375 mm, respectively.

[24] Figure 12 shows the time series of 5-day accumulated evapotranspiration (ET) during the GS. In 1999, ET attained an annual maximum of 1.74 mm/day in mid-July, while ET reached an annual maximum of 1.48 mm/day in mid-August 2000. The ET in 1999 was 0.2–0.4 mm/day higher than in 2000, except during late June and mid-August.

[25] In most applications, differences in lateral flow terms and the rate of downward drainage are negligible when evaporation is the only moisture depletion mechanism in the soil profile [*Brutsaert*, 1982]. The Arvaikheer site has a very flat surface covered by naturally short grass. As described in section 4.1, soil moisture was less than 20%. Therefore we can assume that the downward drainage of water is negligible. Thus the water budget equation for soil layer can be written as

$$P = ET + \Delta W, \tag{4}$$

where P is the rate of precipitation (mm), ET is evapotranspiration (mm), and ΔW is the change of stored moisture in the soil (mm). Change in observed soil water content was used to calculate ΔW. Figure 13 shows the time

Table 2. List of Soil Physical Parameters at the Observation Site

Depth, cm	Porosity, $cm^3\ cm^{-3}$	Dry Density, $g\ cm^{-3}$	Saturated Hydraulic Conductivity, $cm\ s^{-1}$	Soil Type
10	0.475	1.35	2.5×10^{-2}	brown humus sand
35	0.298	1.79	1.83×10^{-1}	brown soil with rock
55	0.335	1.69	2.44×10^{-4}	brown silt

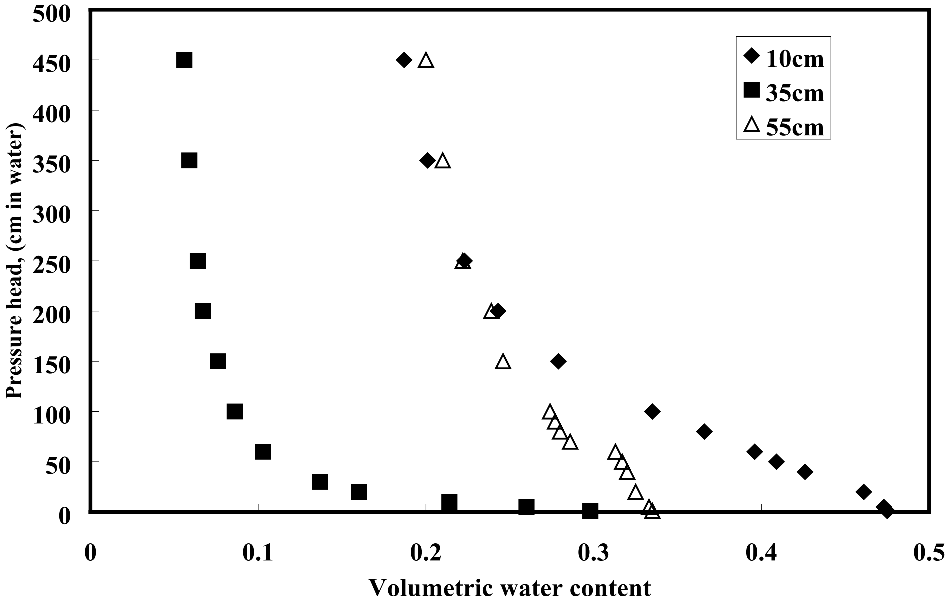

Figure 5. Moisture characteristics curve of soil sampled at Arvaikheer and measured in a laboratory using a suction plate assembly method.

series of 5-day cumulative values for each component (P, ET, and ΔW) of the water budget during the 1999 and 2000 growing seasons, respectively. The cumulative P before July was 96.7 mm and 60.7 mm in 1999 and 2000, respectively. The rain gauge using in Mongolia is Tretyakov type with wind shield and measured at 2 m. The average catch ratios range from 81 to 97% for rain [*Yang et al.*, 1995], which corresponds from 4.5 to 28.5 mm when annual precipitation was 150 mm. Therefore the difference of cumulative P (36 mm) before July was significant. In 1999, cumulative ΔW gradually increased in May and abruptly increased from 4.7 to 20.7 mm following an increase in cumulative P; cumulative ΔW remained nearly zero until the end of July 2000. In mid-July 1999, cumulative ΔW reached an annual maximum of 27.1 mm when θ reached its annual maximum (cf. Figure 6). In mid-August 2000, cumulative ΔW reached an annual maximum of 14.9 mm when θ was at its annual maximum (cf. Figure 6). In 1999, averaged ET/P and ΔW/P during the GS were 0.79 and 0.15, respectively. In 2000, averaged ET/P and ΔW/P during the GS were 0.94 and 0, respectively. The water budget was thus not completely closed. The remaining water, however, is only about 5% of P, which is less than the uncertainty in evaluating evaporation and

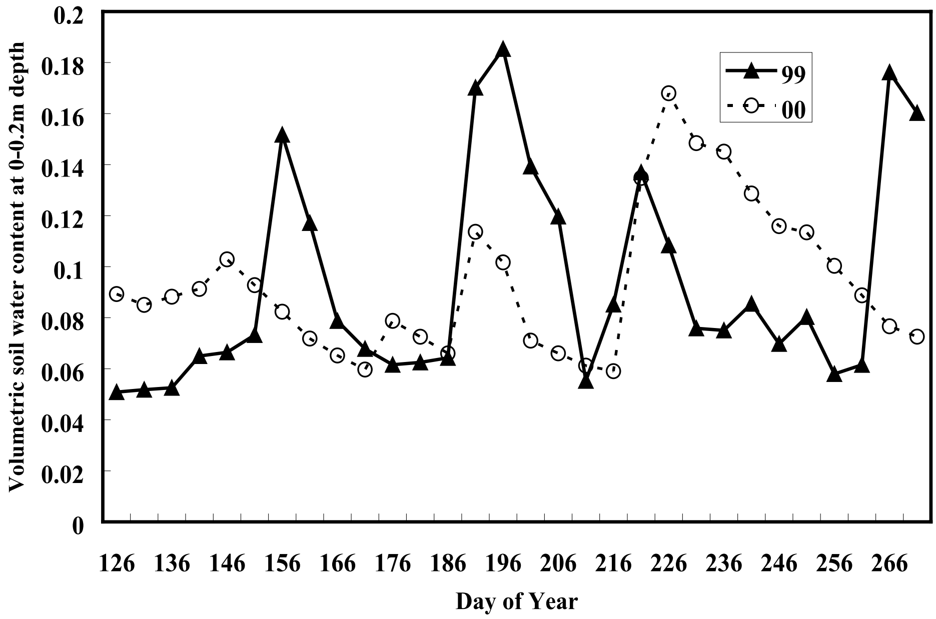

Figure 6. Comparison of 5-day mean volumetric soil water content averaged from 5- to 20-cm depths from May to September in 1999 and 2000.

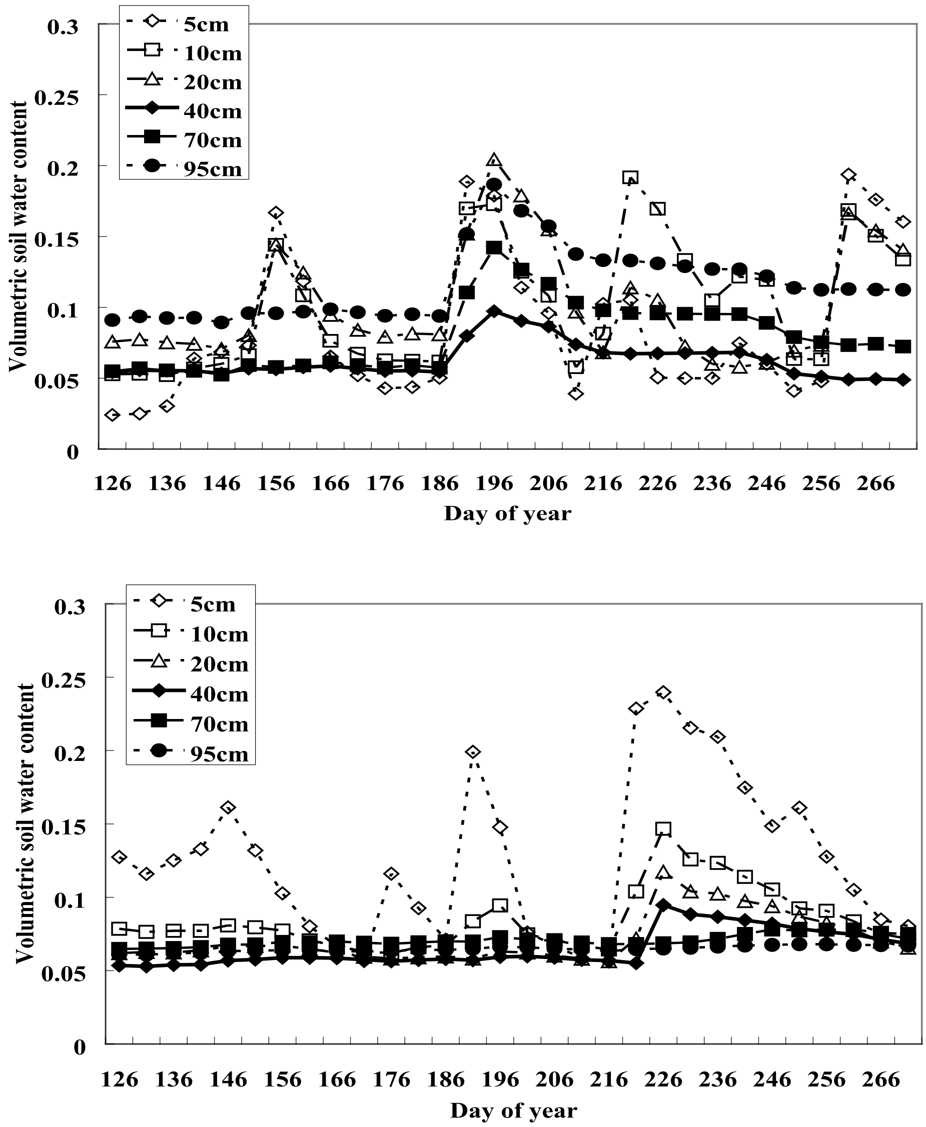

Figure 7. Time series of 5-day mean volumetric soil water content at 5, 10, 20, 40, 70, and 95 cm in (top) 1999 and (bottom) 2000.

measuring P. Results suggest that about 15% of precipitation was stored in the soil layer, and approximately 80% evaporated in 1999, which resembled results from the mean value of the classical estimation of evaporation near Arvaikheer [*Tuvedendorzh and Myagmarzhav*, 1985], and Selenge river basin in Mongolia [*Ma et al.*, 2003]. In contrast, almost all P evaporated and little to no water was stored in the soil in 2000, which is similar to the highest value of Selenge river basin [*Ma et al.*, 2003] and the characteristics of whole Mongolia [*Natsagdorj*, 2000].

5. Summary and Remarks

[26] In this study, we investigated the impacts of seasonal and interannual variability on vegetation as well as soil moisture and evapotranspiration over a grassland at Arvaikheer, central Mongolia. Correlations were determined for 19 years (1982–2000) of monthly mean LAI and climate data. The largest and significant correlation was found for P in July with LAI in July, which continued until September. Furthermore, significant correlation showed for P in May and LAI in June, P in June and LAI in June. In the composite analysis of H5 and L5, the significant positive anomalies of P (45.4 mm) in July and LAI in July (0.6) and August (0.6) were found, which was corresponded to the correlation analysis. It implies that the P in July had the largest impact for grass growth.

[27] To study the physical processes of grass growth, we conducted a case study using energy and water budget data with hydrometeorological elements from 1999 and 2000. In both 1999 and 2000, total precipitation during GS was almost same about 130 mm. However, the intraseasonal distribution of rainfall was different. P in 1999 before July was about 70% of total P during GS while P in 2000 before July was about 40%. These intraseasonal distribution of P caused the difference of the intraseasonal variation of surface θ. The θ in upper 20 cm in June and July 1999 was higher than in 2000. The LAI from June to September

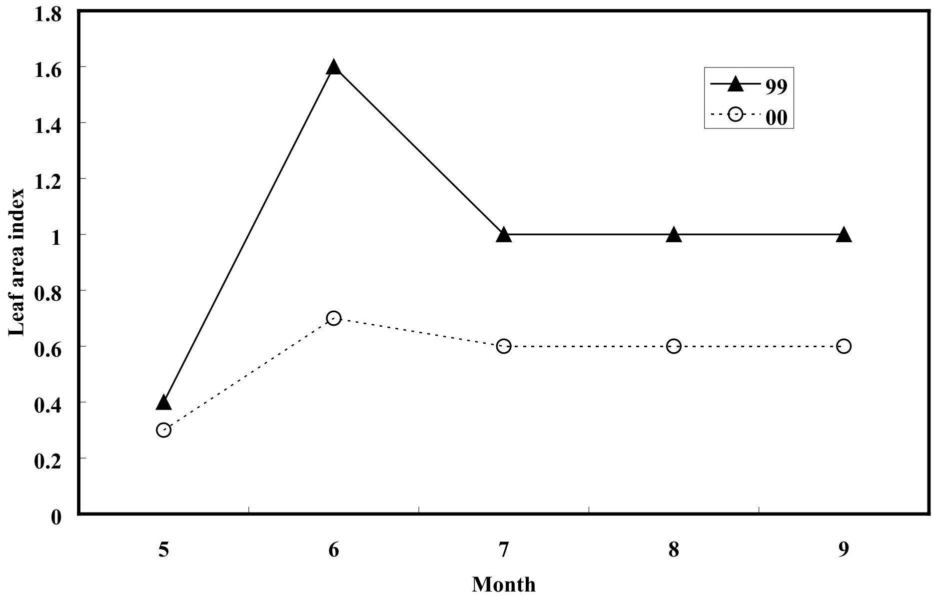

Figure 8. Comparison of the monthly LAI from May to September in 1999 and 2000.

in 1999 was always larger than in 2000. These differences of surface conditions were closely related with the average ET between 1999 (1.3 mm/day) and 2000 (0.7 mm/day). In 1999, averaged ET/P and ΔW/P during the growing season were 0.79 and 0.15, respectively. In contrast, ET/P and ΔW/P during the growing season were 0.94 and 0, respectively. There was almost no stored water in the soil in 2000 because of low P frequency and the small P amounts in June and July 2000.

[28] The results of this study suggest that grass growth in central Mongolia is influenced by P and ΔW before July. High amounts of P, provided after August, did not contribute to grass growth. Although this study focused on one site, it is the first study to examine physical processes related to seasonal rainfall variation and its impact on grass growth in central Mongolia. To generalize our results, we are conducting further analyses using LAI data and data from other stations.

Appendix A: Details About the Observation Done by AWS

[29] Wind velocity and direction were observed at 9.8 m with a wind-vane and anemometer (model: 09101, R.M. Young). Soil heat flux was calculated from values recorded by a heat plate (model: HFT3.1, REBS) buried 0.05 m below the surface. Soil-temperature profiles were obtained by a platinum thermometer (model: STP-1 Pt, REBS) at a 0.05-m depth and a thermistor at 0.1, 0.2, 0.4, 0.7, and 0.95 m depths (model: HOBO Temp, Onset). Volumetric

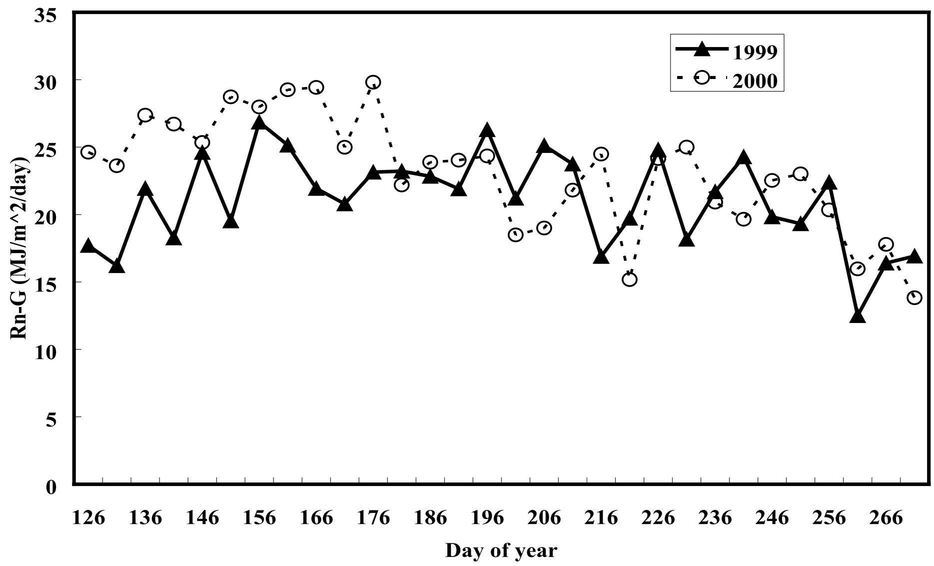

Figure 9. Comparison of 5-day mean available energy values from May to September in 1999 and 2000.

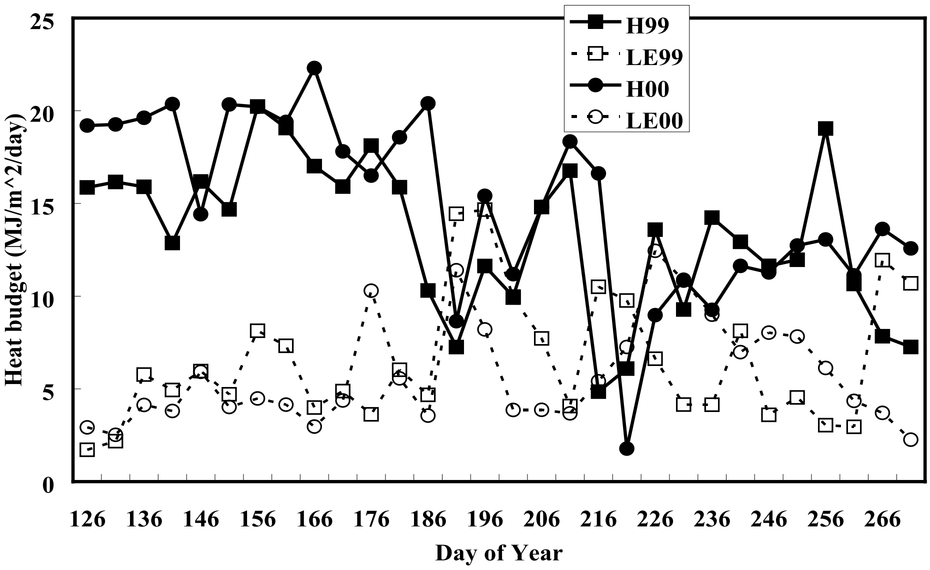

Figure 10. Comparison of 5-day mean sensible heat flux and latent heat flux values from May to September in 1999 and 2000.

soil water content was observed at the same depth as soil temperature by a time domain reflectometry (TDR) sensor (model: TRIME-MUX6, Imko). We used a four-component radiation system at 1.5 m to measure the radiation budget. The system measured shortwave (model: PSP, Eppley) and long-wave radiation (model: PIR, Eppley) in both downward and upward directions; an aspirating fan reduced the effect of dome and case heating. The system also measured net radiation using a net radiometer (model: Q7, REBs) and solar radiation using a pyranometer (model: LI1200SA, Licor). Net radiation was calculated using data obtained from the four-component radiation system; long-wave radiation was corrected using pygeometer dome and case temperatures and a dome coefficient (=3.0), derived from *Albrecht and Cox* [1977].

[30] The PAM III station employs a 3-D sonic anemometer (model: R3A, GILL) and a hygrothermometer (model: 50Y, Vaisala) to determine surface turbulent fluxes of momentum and sensible heat flux by the eddy correlation technique; latent heat flux is determined by the bandpass covariance method, using the Environmental Variable Extractor (EVE; *Horst and Oncley* [1995] and *Horst et al.* [1997]). The bandpass covariance method can serve as an alternative to the eddy correlation method since it does not require ideal sensor response covering a wide frequency range. Rather, the method essentially uses direct measure-

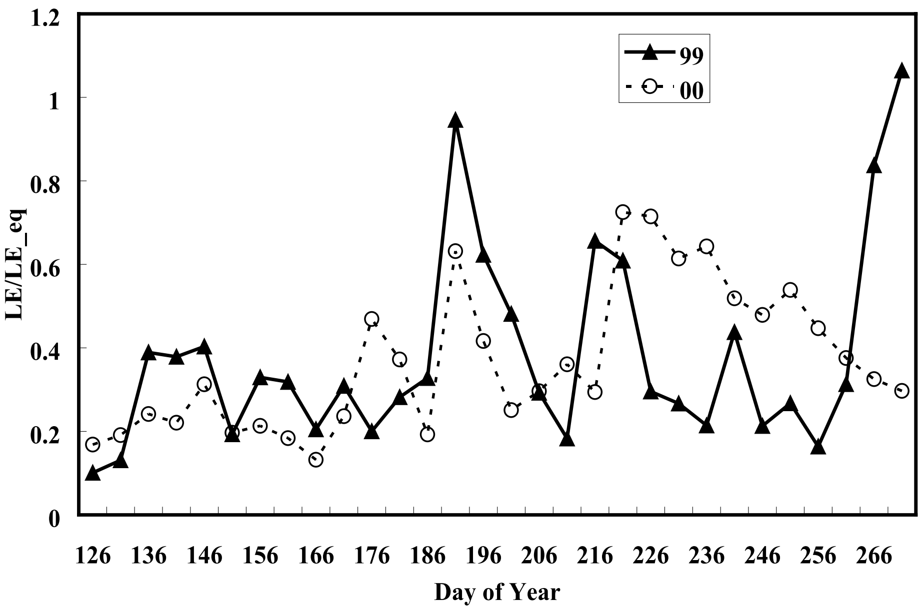

Figure 11. Comparison of 5-day mean LE/LE_{eq} from May to September in 1999 and 2000.

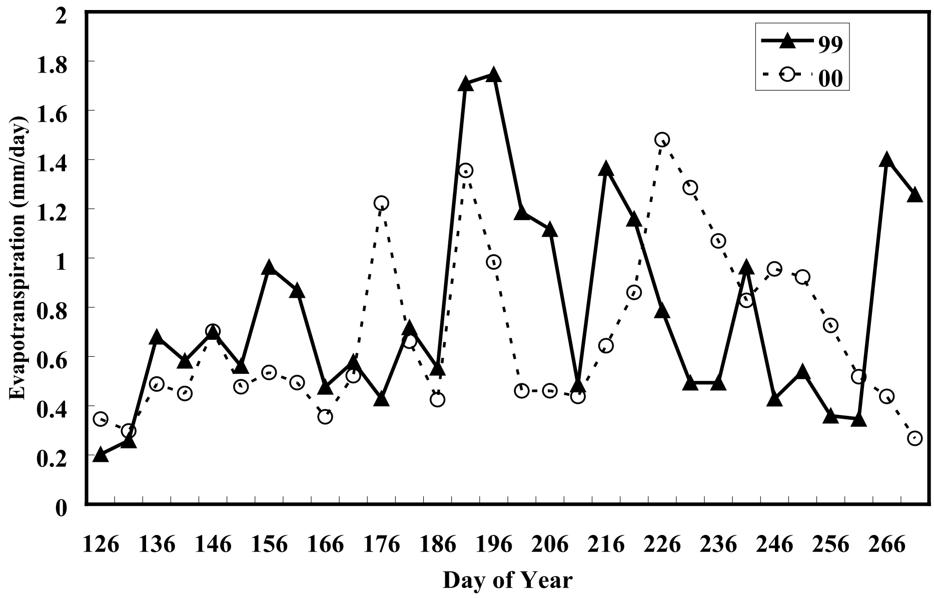

Figure 12. Comparison of 5-day mean evapotranspiration values from May to September in 1999 and 2000.

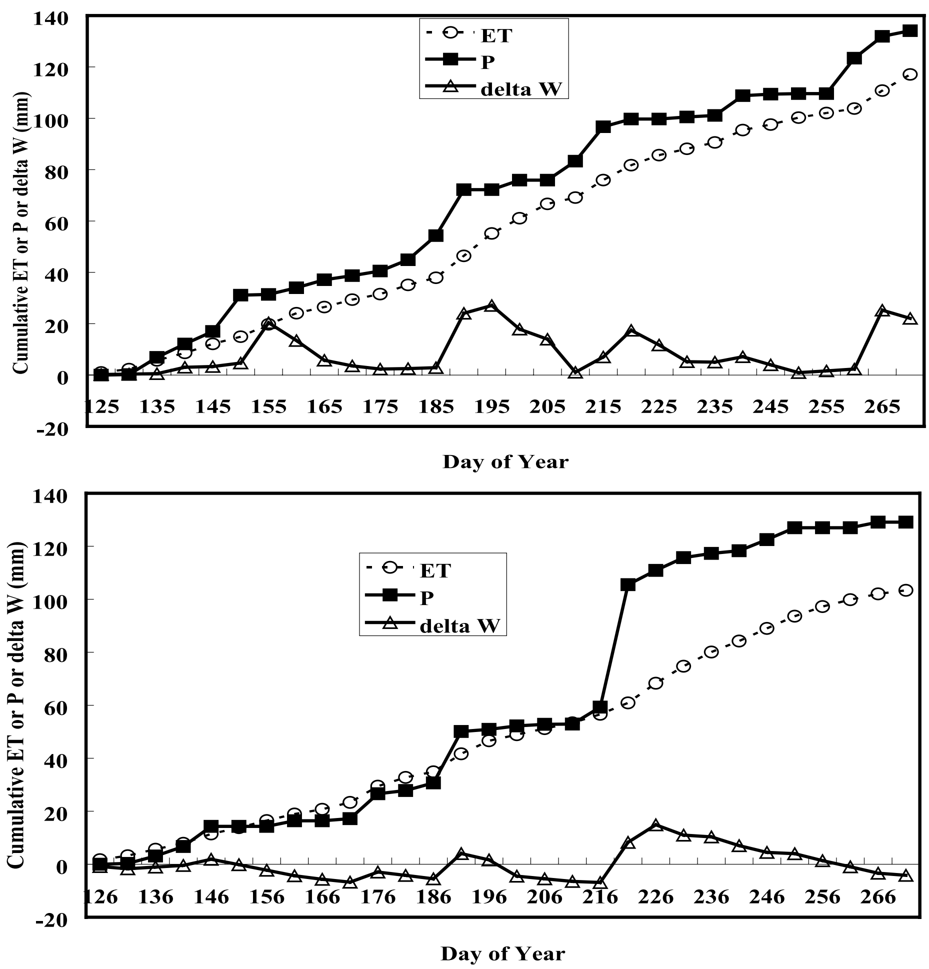

Figure 13. Time series of 5-day cumulative precipitation, evapotranspiration, and change of stored soil moisture values from May to September in (top) 1999 and (bottom) 2000.

ments [*Watanabe et al.*, 2000]. *Watanabe et al.* [2000] suggested that the bandpass covariance method is the best solution for scalar flux in a long-term continuous direct measurement method. *Toda et al.* [2002] provide further details of the bandpass covariance method used here.

[31] Although observations at the study site began in 1997, missing data present a problem until 1998. Therefore we used data from 1999 and 2000. Owing to a lack of nighttime data, we mainly used daytime (i.e., 0900–1600 LST; GMT +8) mean values for this study. Problems with EVE created gaps in the data set for 1999. To fill these gaps we used data from other AWS operated at the same site. AWS was made by AANDERA Instruments, Norway, and have 3.6-m tower heights. They consist of a pyranometer (model: 2770, AANDERA), thermohygrometer (model: HMP35D, Vaisala), anemometer (model: 2740, AANDERA), wind vane (model: 3590, AANDERA), barometer (model: 2810, AANDERA), and soil moisture sensor (model: CS615, Campbell). Meteorological sensors are 3.6 m high and soil moisture sensors are positioned at depths of 5, 10, 20, 40, 70, and 95 cm. Time intervals for the Aandera weather stations were 10 minutes or 30 minutes, and the data collection period was from February 1999 to June 2000. Before gap filling, comparison between AANDERA and PAM III data sets was carried out for a period without any gaps of PAM III data.

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