Limiting Factors for Nomadic Pastoralism in Mongolian Steppe: A Hydrologic Perspective

Michiaki Sugita[1*], Shintaroh Yoshizawa[2**], and Ishgaldam Byambakhuu[2***]

[1] Faculty of Life and Environmental Sciences, University of Tsukuba, Ibaraki 305-8572, Japan
[2] Graduate School of Life and Environmental Sciences, University of Tsukuba, Ibaraki 305-8572, Japan

[*] Corresponding author, sugita@geoenv.tsukuba.ac.jp
[**] Now at Graduate School of Comprehensive Human Sciences, University of Tsukuba, Ibaraki, Japan
[***] Now at Department of Environmental Sciences and Chemical Engineering, School of Engineering and Applied Sciences, National University of Mongolia, Ulaanbaatar, Mongolia

---

The final publication is available at ScienceDirect via http://dx.doi.org/ 10.1016/j.jhydrol.2015.02.050

Abstract

In this study, limiting factors for continuing nomadic pastoralism in steppe areas were studied based on a hydrologic perspective. Two small watersheds in central Mongolia were selected for an assessment of water balance and hydrologic processes. We determined that the majority of annual precipitation, ~88-96 mm, was lost by evaporation (82%) while only a small proportion went to groundwater discharge, surface runoff, and groundwater consumption by nomadic activities. The soil column was found to absorb large fluctuations in precipitation although its connection to groundwater was very weak. Groundwater recharge was, therefore, very small and occurred only rarely during heavy rainfall events in valley bottoms. However, current water storage in shallow groundwater was determined to be quite sufficient for continuing nomadic pastoralism when compared to the drinking water requirements of livestock. The main limiting factors identified were a temporal lack of feed to animals due to a loss of aboveground biomass resulting from soil moisture shortages during drought conditions, and a decline in the number and maintenance level of the traditional well network that, due to access to shallow groundwater, has allowed herders to migrate to areas with better conditions in remote Mongolian steppe.

Keywords: water balance; groundwater recharge, nomadic pastoralism; steppe; Mongolia

1. Introduction

In remote rangelands, water is often considered to be a limiting factor for continued nomadic pastoralism because drinking water limitations influence the sustainability of livestock. In most cases, with the exception of camels which can survive for extended periods without drinking water, livestock must consume water on a daily basis. Since perennial streams are rare to non-existent in many arid regions, daily water consumption largely depends, with the exception of regions near mountains where spring water is expected and the winter season when accumulated snow and ice serve as a water, on shallow groundwater.

Since nomadic pastoralism has continued over centuries (e.g., Miyawaki, 2002; Matsubara, 2005; Cribb, 1991), implying sustainability, the impact of livestock on shallow groundwater stocks should not be very large. Crude estimates of annual water balance support this assessment. For example, by assuming an annual rainfall value of $P = 150$ mm, an evaporation value of $E = 0.8 \times P = 120$ mm, a daily water consumption value for sheep of 4 L, and a typical stocking rate of 0.5 head/ha, an annual sheep water consumption of 730 L/ha and an annual available water value ($=P$-E) of 3.0 10^5 L/ha can be derived. Given these values, water consumption by livestock only constitutes 0.2% of available water. However, such an argument is based on average conditions. A relevant question is whether or not water is sufficiently available everywhere and at all times, and if not, what could induce a temporal/local water shortage. Additional questions include the following: (i) Is water abundant in every season of the year? (ii) Is there enough water even under drought conditions when annual rainfall is much less than the climatic mean value? (iii) Is the stocking rate as low as the assumed value everywhere? (iv) Will the stocking rate continue to be at the same level as the assumed value? (v) Were stocking rates and water balance regimes different in the past? Such questions address some of the relevant factors that could influence the sustainability of nomadic pastoralism. To answer such questions, crude estimations are not sufficient. A comprehensive study with more reliable data and statistics is clearly required.

The above discussion presents a brief background for the present study. We performed a detailed water balance and estimated hydrologic processes for pastureland in order to shed some light on the limiting factors for nomadic pastoralism. Additionally, vegetation dynamics in relation to the water deficit were considered by referring to information presented in the companion study of Satoh et al. (2012), and were based on detailed ecohydrologic observations and the carbon dynamic simulation results of Chen et al. (2007), among others. Since a lack of water can also limit nomadic pastoralism due to the influence of water availability on vegetation growth, information on vegetation dynamics was also obtained. Combining water balance estimates and ecohydrological information allowed us to answer the question of why nomadic pastoralism has been able to continue over centuries under arid conditions.

When a water balance approach is applied for estimating minor hydrologic components such as groundwater recharge rates, the resulting estimates tend to involve a large error (e.g., Healy, 2010). Therefore, our estimations were not only based on the water balance method but also on supplementary observations. Analyses were also applied whenever possible in order to enhance the reliability of water balance estimates. Two small watersheds located in the central portion of Mongolia were selected for our analyses. Therefore, the work presented here is essentially a case study. However, Mongolian steppe is a part of the largest rangeland belt on Earth, extending from northeastern to Central Asia (Shiirevdamba, 1998), thus, our results should be applicable to larger areas containing similar vegetation and topography.

2. Method

2-1. Study area

For the study, two small watersheds, W1 and W2, were selected within a gentle hilly terrain in Saintsagaan soum (county) in Dundgobi aimag (prefecture), located approximately 4 km northeast of the city of Mandalgobi in Mongolia (Fig.1). The watershed area of W1 is A = 3.3967 km^2 and that of W2 is A = 11.3567 km^2 (see the Appendix for watershed boundary determinations). The soil in this area is classified as semidesert brown soil or semidesert brown friable sandy soil (Dorzhgotov 2003). The morphology of the A horizon can be characterized by high compactness, a brown color, and weak development of the soil structure. A high sand and gravel content can also be observed within the surface soil (Asano 2010, personal communication). The bedrock is composed mainly of Permian felsic rocks such as dacite, rhyolite, or quartz porphyry (Mineral Resources Authority of Mongolia, 2001; Mineral Resources Authority of Mongolia and Mongolian Academy of Sciences, 1998). The climate is classified as arid and has a mean annual precipitation of 150 mm (for 1944–2011), as measured by the nearby meteorological station (located approximately 5.5 km southeast from the study area) at the Research and Information Institute of Meteorology, Hydrology and Environment (hereafter referred to as the IMH station). According to the Standardized Precipitation Index (SPI) for three, six and 12 months, the observation period from 2008-2011 can be classified as “near normal” to “moderately dry” (WMO, 2012). Approximately 78% of annual precipitation falls during the summer from June-August. The mean annual air temperature is 1.5°C, while the annual range of monthly means is as large as 37°C, with a low of -17.7°C in January and a high of 19.3°C in July.

According to the Mongolian vegetation classification scheme, the area is located within a boundary zone between the desert steppe class and the steppe/dry steppe class. Dominant vegetation consists of the C_3 herbaceous plant *Allium polyrrhizum* and some dotted communities of the C_3 shrub *Caragana microphylla*. However, during the study period, even during the typical growing season from May through August, only intermittent appearances of *Allium* were observed at the surface after rainfall events (also see Satoh et al., 2012). Plants in this area, especially *Allium*, are under grazing pressure mainly from the sheep and goats of local nomadic herdsmen (Byambakhuu et al., 2010, Satoh et al, 2012). The mean number of animals per unit area (i.e., the grazing pressure or the stocking rate) in 2009 was $S_r = 0.79$ SEU ha^{-1} for Saintsagaan soum (3.39×10^3 km^2) and $S_r = 0.35$ SEU ha^{-1} for Dundgobi aimag (7.47×10^4 km^2), while, as determined from local animal population numbers (see below), it was approximately $S_r = 0.56$ SEU ha^{-1} within watersheds W1 and W2. SEU is the sheep equivalent unit and represents the population sum for five major animal types in Mongolia (sheep, goats, horses, cattle and cows, and camels) converted to sheep equivalent numbers using common factors (e.g., Pratt and Rasmussen, 2001). Thus, the average stocking rate for the two watersheds is between the higher S_r of Saintsagaan soum and the lower value of Dundgobi aimag, implying that the target study areas reflects a somewhat modified herding condition due to the proximity to the city of Mandalgobi. However, conditions within the W1 and W2 watersheds represent more of the conditions of rural rangelands in Mongolia than those in central portions near its capital city of Ulaanbaatar where a $Sr \geq 1.25$ is often observed (see e.g., Fig. 12 of Sugita et al., 2007).

In and around the W1 and W2 watersheds, there are four wells (GW1-GW4) completed in shallow groundwater (Fig.1) and were used regularly by one to two nomad families, and infrequently by other families, during the observation period. A ‘ger’ (a nomad tent) was placed near GW1 by one family during the summer of 2009. Another family camped throughout the year near GW2. These types of wells are used to obtain access to shallow groundwater and are the main water resource used by local herdsmen for daily life. More specifically, the GW1 well (with a depth of 5.5 m; where the typical water depth is 0.5 m) was only used regularly during the summer season of 2009 by one family, while from 2010 it was only periodically visited and used by nomads from the W2 watershed. Wells GW2 (with a depth of 3.1 m and a water depth of 0.5 m) and GW4 (with a depth of 1.7 m and a water depth of 0.2 m) were used throughout the year by this family. The GW3 well (with a depth of 2.1 m and a water depth of 0.1 m) did not provide sufficient water and was not normally used by the nomad families.

The surface located 50-100 m from the wells and the 'ger' was essentially bare ground without vegetation, a result of micro-scale overgrazing. Such concentrated overgrazing is common in this area of Mongolia but generally disappears or decreases exponentially from 50-100 m as distance from the source point of grazing, such as a well, increases (e.g., Sasaki et al., 2008). Therefore, the extent of such bare ground is only limited to areas surrounding wells and 'gers' and, thus, impacts to watershed-scale water balances and hydrologic processes can likely be ignored.

2-2. Estimations of water balance components

The water balance of a watershed for a given period can be expressed with a dimension of [LT^{-1}], as follows:

$$P = E + G_{out} + R_{out} + U + \Delta S \tag{1}$$

where P is precipitation, E is evaporation (soil evaporation and transpiration), G_{out} is groundwater discharge from the watershed, R_{out} is surface runoff from the watershed, U is the amount of groundwater extracted through the wells used mainly for grazing, and ΔS is the storage change within the watershed over the period in which (1) is determined. In the present study, (1) was considered from the surface down to the bottom of the shallow, unconfined aquifer and we assumed that the shallow aquifer is underlain by an impermeable layer so no leakage occurs. Since the observed groundwater level did not show steady declines during the dry period (see the Results section and the Appendix), such an assumption is likely acceptable.

Although the exact thickness of the shallow aquifer is unknown, the Well Database of Mongolia, currently being compiled by the IMH (Davaa, G., personal communication, 2008), lists the thickness recorded for each well in Mongolia. The compilation process allowed us to estimate an average (of 5.1 m) for the 15 wells registered in and around the W1 and W2 watersheds.

For estimations of each term in equation (1), two sets of automatic weather stations (AWS), MG1 (45°49'11.30"N, 106°17'43.00"E) and MG2 (45°48'34.80"N, 106°16'51.40"E), were installed on 7 July 2008 and 3 May 2009, respectively, in order to measure the basic meteorological and hydrologic variables (Fig. 1 and Table 1) required for estimating evaporation and soil moisture storage. The MG1 station is located within the *Caragana* community while the MG2 station is located on a surface predominantly covered by *Allium*. An extensive and wide range of hydrologic elements were also obtained. From the data, each term for (1) was derived directly or estimated using an appropriate theory and methodology. Details are provided in the Appendix. Briefly, the estimation/measurement schemes can be summarized as follows. Precipitation, P, was derived from measurements obtained from the MG1 and IMH stations. Estimates for evaporation, E, were made by applying the energy balance method, using sensible heat flux estimated from the locally calibrated bulk equation. Groundwater discharge, G_{out}, and surface runoff, R_{out}, from the watersheds were derived using a combination of topographic surveys, groundwater level measurements, well pumping tests, and information provided by local headsmen, while the amount of groundwater extracted through the wells, U, was estimated based on animal numbers and the daily unit of water consumption for each animal (Table 2 and Appendix). Storage change, ΔS, within the watershed was estimated from groundwater level observations and soil moisture measurements obtained from the AWSs by assuming that these observations represented average conditions for the watersheds.

3. Results

Our observations and analyses are summarized in Figure 2 (seasonal changes of water balance), Table 3 (annual water balance), Table 4 ($P-\Delta S$ and mean residence time for different soil depth ranges), and Table 5 (the mean residence time $\bar{t}$ for shallow groundwater). A discussion of our results is found below.

3.1 The distribution of water at the surface: Where does precipitation go?

On an annual basis, the answer to the question above is quite clear - the majority of precipitation, P, is lost to the atmosphere by evaporation, E. However, the E/P ratio was found to greatly change both seasonally and annually. The other terms of water balance were very small and may actually be neglected depending on the purpose of the analysis.

During the three, one-year periods for which annual water balance was estimated (Table 3), the annual precipitation value was 87.5-95.8 mm and corresponded to 58-64% of long-term mean annual precipitation (150 mm) for Mandalgobi. As mentioned in the Methods section, the observation period was characterized by weak drought conditions. Seasonally, as can be seen from Fig. 2, inputs to the watershed occurred as rainfall almost entirely during the warm season. Winter precipitation (i.e., snowfall) was minor and totals were 3.9, 6.3, and 8.0 mm for the 2008-2009, 2009-2010, and 2010-2011 winter seasons, respectively.

In contrast to precipitation, evaporation took place not only during the warm season but also during the winter season. For example, total evaporation at MG1 from December 2008 through February 2009 was 14.1 mm. The result is somewhat surprising since Mongolian winter is known by its severely cold climate, during which one would expect little evaporation. However, the winter climate is also characterized by continuously sunny weather and by a persistently cold Siberian air mass. As a result, Penman's potential evaporation, *PE*, during the winter can be as large as 87.5 mm, approximately 8% of annual *PE*.

As can be seen from Table 3, annual evaporation was 65.1-83.7 mm, representing ~68-96% of precipitation during the same period. The value is close to that reported by Yamanaka et al. (2007), obtained from measurements at four sites surrounding the Mandalgobi area during the summer season from June through August. However, our results indicate that this ratio changes markedly depending on the target period. To understand the availability of energy and water for evaporation, these ratios may also be compared with standard relationships such as the Budyko or Turc-Pike functions and E/P and PE/P (e.g., Arora, 2002). Based on Table 3, the 3-year averages can immediately be obtained as E/P = 0.82 and PE/P = 11.6. In a figure to show functions with PE/P on the *x*-axis and E/P on the *y*-axis, the values are plotted within the upper-right corner of the figure (e.g., Fig.1 of Aurora, 2002). General agreement exists although E/P = 0.82 is located on the lower side expected for the high aridity index of PE/P = 11.6.

Overall, annual water balance is dominated by precipitation and evaporation, and to some extent by infiltration into the soil layer (see the discussion below on this topic). As is clear from data provided in Table 3, groundwater discharge (G_{out}), surface runoff (R_{out}), and groundwater use (U) are all minor components of water balance within the two watersheds. It is likely that a small G_{out} and R_{out} would be similar for other watersheds in the area, since the hydraulic gradient that determines G_{out} and the surface gradient that determines R_{out} are expected to be small within this generally flat terrain. With all water balance components either measured or estimated, the water balance (1) should, in theory, be closed. However, in reality this is not the case and the residual (i.e., the error term) is shown as ε in Table 3. On an annual basis, this value is approximately in the range of -24 to 34 mm a^{-1} but is reduced to 3 to 4 mm a^{-1} when the average of three years is considered, suggesting a random, rather than a systematic, error for water balance estimations.

3.2 The role of soil moisture as a cushion for absorbing precipitation fluctuations

The soil column was found to play a major role in smoothing out fluctuating precipitation inputs to each watershed. Such a result is essentially accomplished because the soil column acts as a reservoir. When a drought period continues, soil moisture is released for soil evaporation and transpiration and the growth and survival of vegetation is maintained, while it is replenished during a rainy period. As a result, the availability of water for vegetation fluctuates less than precipitation.

As mentioned above, on average, 82% of precipitation is lost by evaporation each year. The remaining water, 12.7 mm a^{-1} (14% of P), goes to infiltration and contributes to the storage change, ΔS. As can be seen in Fig. 2, on shorter time scales this ratio is also quite variable, indicating seasonal variations for P and SWC and the accumulated value of P and $P-\Delta S$. The value for ΔS in this case was determined based on SWC data from AWSs located at the surface down to -1.3 m. As can be seen from (1), the result should be the same as E under conditions of negligible G_{out}, R_{out}, U, and ΔS within soils below the lowest levels. Thus, the general agreement found in Fig. 2 between the two provides partial verification for the evaporation estimates obtained by (1).

As can be seen from the figure, P occurs intermittently while E and $P-\Delta S$ are continuous, indicating a direct interaction between P and SWC during the wet period and between E and SWC during the dry period. Also clear is that there are periods where $\sum E > \sum P$. For example, beginning in February of 2011 through July of the same year, accumulated evaporation continued to be larger than accumulated rainfall. During these periods, evaporation consumes the stored soil moisture and drought conditions can easily result. During the observation period, however, this type of outcome did not occur for extended periods.

Some finer points in Fig. 2 also require clarification. First, during the winter period, in contrast to the value of E obtained from the AWS, $\sum(P-\Delta S)$ did not display an increase. The result is not surprising since TDR sensors are incapable of detecting accurate soil moisture when the soil is frozen (see the Appendix for more information). For this reason, the ΔS values estimated during cold seasons should be treated with care. The second point to notice in Fig. 2 is two occasions of decrease for $\sum(P-\Delta S)$. For the spring of 2010, as indicated by an increase in SWC and soil temperatures above freezing, the decrease was due to the infiltration of melted snowfall. Melted snowfall is not included in the calculation of P and thus resulted in a decrease for $\sum(P-\Delta S)$. For spring 2009, however, the infiltration of melted snowfall had already occurred by the end of March and cannot explain the decrease of $\sum(P-\Delta S)$. One possible explanation for the underestimation of P is a tipping bucket rain gauge with accumulated sands due to the dust storms that typically occur in the area during March-April.

3.3 Soil water-groundwater interactions

We determined that the groundwater-soil moisture connection was very weak during rainfall events and on rain-free days. As a result, groundwater recharge rarely took place and only occurred during heavy rainfall events mainly in valley bottoms. Groundwater is also not directly used by transpiration. Thus, the role of groundwater in the hydrologic cycle is quite limited, as can be confirmed by the observational results and the analyses explained below.

Table 4 lists the values of $P-\Delta S$ estimated for various depth ranges. The list indicates which soil layers are the most actively involved in the annual hydrologic cycle within the area. Clearly, the top soil layers play a major role. $P-\Delta S$ values for the layer from 0-0.25 m are roughly 85-90% of those for the maximum depth range of 0-1.3 m. Even those for the 0-0.075 layer accounted for some 70% and can also be compared with the stable groundwater levels of GW1, GW2, and GW4 (Fig.2) that usually do not respond to rainfall events (see the discussion below).

Table 4 also lists the mean residence time of soil water which was estimated by assuming a steady state condition for each depth range. For the most active layers of 0-0.25 m, this is on the order of months, while for the range involving deeper layers it is on the order of a year. Table 5 provides the mean residence time, $\bar{t}$, for shallow groundwater as estimated from an analysis of the tritium concentration, C, of well water samples taken on 12 May 2011 (see the Appendix for more details). The value of $\bar{t}$ was found to be in the range of 30-60 years. Due to a limited number of available groundwater samples and the uncertainty involved in the choice of a piston-type flow model, the residence time determined is, thus, not very accurate, but it is clearly much longer than that of soil water, in agreement with results derived from the water balance consideration outlined above. The larger picture here is that the shallow groundwater-soil water connection is very weak to non-existent and periodic connections in the form of groundwater recharge take place only rarely during heavy rainfall events.

The arguments outlined above, based on mean residence time, can also be supplemented by looking at the time-series data. For example, an apparent decrease in groundwater levels due to evaporation was not observed (Fig.2), indicating a limited connection between groundwater and soil water during sunny periods. The result is in agreement with the findings of Satoh et al. (2012) who indicated, using stable isotope analyses, that within the W1 and W2 watershed groundwater is not directly used for transpiration.

During rainfall events, it is obvious that not all SWC values throughout the soil column responded to all rainfall events. In general, only SWC values at shallower depths responded, while those at deeper layers of 70-110 cm did not display an obvious response. However, three cases were determined when all of the SWC values increased, as follows: (i) at the end of April 2009, (ii) during the March-April period of 2010, and (iii) at the end of June 2011. The increase in SWC at the end of August 2008 may also be added to this category, although it is not quite clear if the change in SWC at 70 cm and 110 cm were due to target rainfall events or to events that took place earlier. Among them (including the questionable August 2008 case) only two cases occurred in which an increase in groundwater level was also observed. One was for the rainfall event on 29 August 2008 (total rainfall of 23.5 mm and a maximum intensity of 0.7 mm 10 min^{-1}) for which a small water level increase of GW2 was observed. Another one occurred on 22 June 2011 (total rainfall of 18.5 mm and a maximum intensity of 6.3 mm 10 min^{-1}) when a GW4 water level increase was observed. These events are, thus, rare occasions when groundwater recharge took place. However, by comparing SWC and groundwater level changes, it is obvious that the increases in groundwater level at GW2 and GW4 were more pulse-like than gradual and took place much earlier than those of the deeper SWC at 70-110 cm where only slow and gradual increases were observed. The result tends to support an argument that infiltration and SWC increases on a hillslope are not directly connected to the groundwater level increase observed within the valley bottom. Thus, groundwater recharge is not uniform throughout the watershed but, rather, is a localized phenomenon. This type of focused recharge also appears to be more common than distributed recharge in the arid climate (e.g., Scanlon et al., 2006).

Heavy rainfall events and groundwater level increases could also have been accompanied by surface runoff generation in valley bottoms. Although a direct confirmation is not possible, interviews and an analysis of topographic surveys (see the

Appendix) indicate that such an assumption it is not inconceivable. According to our analysis, surface runoff can be expected only once in a 2-3 year period and when rainfall intensity exceeds 4-5 mm within a ten-minute period. For the 22 June event mentioned above, this criterion was satisfied, as graphically shown in Fig. 2. The rainfall intensity of the questionable 29 August event was, on the other hand, too small to have caused surface runoff.

3.4 Groundwater recharge rates

The above argument tends to imply that the amount of groundwater recharge is very small. Using available data, a crude estimation for the mean recharge rate, $\overline{q_z}$, can be calculated using two approaches. For the first approach, (1) should be applied for long-term mean values with $\Delta S = 0$ and $G_{out} = 0$, namely in the form of the following equation:

$$\overline{q_z} = \overline{P} - \overline{E} - \overline{R_{out}} \quad (2)$$

With a climatic mean of $\overline{P}$ = 150 mm a^{-1} (for 1944–2011), an $\overline{E}$ of 82% from $\overline{P}$, and the $\overline{R_{out}}$ obtained from Table 3, (2) leads to $\overline{q_z}$ = 27 mm a^{-1}.

In the second approach, $\overline{q_z}$ was determined from the mean residence time, $\bar{t}$, determined from the tritium concentration analysis and the storage of shallow groundwater, $\overline{S}$, estimated from the mean thickness and the effective porosity of the aquifer (see the Appendix for method details). The estimated value of $\overline{S}$ = 420 mm for both the W1 and W2 watersheds and $\bar{t}$ = 28-59 years yielded a $\overline{q_z}$ in the range of 7-15 mm a^{-1} (Table 5). The values are on a lower side of those estimated by (2), but on the same order of magnitude. Thus, the mean groundwater recharge rate in this region appears to be on the order of 15 mm a^{-1}. Although for this region no additional studies are available for comparing the calculated $\overline{q_z}$, the values are in the range of reported groundwater recharge rates for arid and semiarid regions (e.g., Scanlon et al., 2006), but on the higher side for annual precipitation, $\overline{P}$ = 150 mm.

4. Discussion

4.1 How much groundwater can be used for nomadic pastoralism?

The question above can be addressed by considering that shallow groundwater withdrawal for nomadic pastoralism should not reduce groundwater resources over a long time period. Thus, the maximum amount of groundwater use, $U_{\max}$, should be the same as $\overline{q_z} - \overline{G_{out}}$; determined to be on the order of 15 mm a^{-1}, or 5.1×10^{-5} and 1.7×10^{-4} km^3 a^{-1}, or 1.4×10^5 and 4.6×10^5 L d^{-1} for watersheds W1 and W2, respectively. The result can be compared with an actual water consumption of 9.0×10^2 L d^{-1} (W1) and $3.0\text{-}3.7 \times 10^3$ L d^{-1} (W2), as determined from Table 2. Therefore, it is clear that the level of $U_{\max}$ is sufficient to allow the current level of nomadic pastoralism without a reduction in groundwater resources within this region.

Interestingly, the $U_{\max}$ values are on the same order of magnitude as those preliminarily estimated for four watersheds within northeastern Mongolia where the climate is more humid, has a larger mean annual $\overline{P}$ of 200-250 mm, and a Mongolian vegetation class of mountain forest steppe (Tsujimura, 2007). Since the amount of water is judged to be sufficient for the two extreme vegetation classes, mountain forest steppe and desert steppe, expected for nomadic pastoralism within steppe regions of Mongolia, the result suggests that water is not a limiting factor, in general, for nomadic pastoralism in Mongolia, at least for current climate conditions.

The above discussion provides a long-term mean and is useful for arguing a case for short-term drought. Under a short-term drought, groundwater continues to be used even though no groundwater recharge takes places. Thus, the question is how long current groundwater storage can last in support of nomadic pastoralism. The value can be estimated by deriving a S/U value. With available information for both S and U in watersheds, S/U was determined to be on the order of 10^4 years. Therefore, even during prolonged droughts, shallow groundwater resources are sufficient to maintain current livestock numbers.

4.2 The past and future: Changes in animal numbers, climate, and well networks

The discussion provided above indicates that, at least at present, drinking water from shallow groundwater is not a limiting factor for nomadic pastoralism. The question to consider now is how factors have changed from the past and how they will differ in the future based on current processes? By considering each factor, we determined the crucial factor required for sustainable nomadic pastoralism in Mongolian steppe. As explained below, the presence of a well network in Mongolian steppe that guarantees access to shallow groundwater is the crucial factor.

Animal numbers

The present animal population in Mongolia is the largest it has ever been, particularly since the introduction of a market-oriented economy in 1990-91. In 2007, the total animal population in Mongolia was 40×10^6, and can be compared with much smaller and more stable numbers of 24×10^6 in the 1930s and 26×10^6 in 1990 (see Fig.2 in Sugita et al, 2007). In Saintsagaan soum, a steady increase in livestock number following the introduction of a market-oriented economy in 1990-91 was observed (Fig. 3), with the exception of the 1999-2000 period when a nationwide 'dzud' (a disaster with damage and death to livestock caused by severe cold weather) took place. Thus, in the past, the impact of grazing as compared to water balance should have been even smaller. Considering possible impacts under assumed cases of increased grazing pressure for the future is also possible. For these estimates, the U value was determined for animal number increases that were 200%, 300% and 1000% of current values. Our result indicated, however, that U remained a minor component of (1), even with these dramatic increases. In reality, although the population of livestock increased significantly after 1990-91, their numbers did not increase by more than 100%.

Climate

The past climate can be estimated, to some extent, by looking into records for historical precipitation, reconstructed based on tree-ring analyses. For example, Pederson et al. (2001) provided a reconstructed precipitation record for the period from 1651-1995 for the northern portion of Mongolia, while Jacoby et al. (1999) provided one for central Mongolia. Both studies indicate that there were no apparent continuous trends of increase or decrease in precipitation. Pederson et al. (2001) also described the presence of cyclic short-term wet and dry periods, with a standard deviation of 50.2 mm, and a significant peak in the power spectra at 10.8 and 12.8 years. Based on instrument observations, the results were found to be similar to those obtained for the recent period from 1942-1995. Thus, it is not unreasonable to assume that long-term means and fluctuations of climate have remained approximately the same. Thus, the statement that drinking water from shallow groundwater is not a limiting factor for nomadic pastoralism in Mongolia can be extended to a longer time period. Such a result is also in agreement with the study of Brutsaert and Sugita (2008) who found no clear trend in long-term groundwater storage within the Kherlen River Basin (a target area slightly different from the present study) in northeastern Mongolia from

1947-2006.

Larger uncertainties exist regarding future climates. However, projected future climate studies are available. For example, using two dynamic downscaling schemes in general circulation model outputs (based on the ‘A2’ scenario) using a regional climate model, Sato et al. (2007) indicated a general decrease for precipitation on the order of 20 mm during the July-August period for 2070s Mongolia as compared to present. Since the present recharage rate is very small anyway, the level of decrease should likely only have a minor impact on groundwater. However, for vegetation growth, this change could have a strong impact and is, therefore, discussed below.

Well network

Another factor that must be considered for the past and future is the presence of wells. As mentioned above, when livestock migrates with herders, both periodically and seasonally (e.g., Cribb, 1991), seeking a continuous supply of water and a pasture for grazing, the main water source is shallow groundwater. For water to be available, the widespread access to shallow groundwater via the presence of wells is required. Without shallow wells animals have no way to make use of water. They also cannot migrate even if sufficient water is available within a shallow aquifer.

In Mongolia, the well network in remote pasturelands appears to have begun during the era of Ugedei Khan (1206-27), who ordered wells to be dug to supply water; listed as one of the four deeds in Mongolian history (Anonymous, 1982; Baranchuluun et al., 2004). Although it is not clear how widespread the initial construction of wells was in the past, Sharkhuu (1975) provides a hint in archived records regarding the various disputes amongst people living in the Qing Dynasty (1616-1912) which extended its territory into present-day Mongolia. Amongst the archived records are conflicts regarding the use of wells. Thus, from these descriptions, it appears that at least in the Qing Dynasty era the presence and use of wells in pasturelands were already common. Such wells continue to exist and were maintained until recently through a social institution called ‘neg nutgiinhan’ before the 1930s (Chuluun and Ojima, 2002) and a pastoral cooperatives system known as ‘negdel’ since the late 1950s.

Following introduction of the market-oriented economy in 1990-91 in Mongolia, ‘negdel’ was privatized and the maintenance function of the wells was abandoned or weakened. As a result, in 2000 the nationwide number of wells in pasturelands decreased to 57% of that found in 1990 (Fig.3). Since 2000, a decrease in well decline slowed and an increase was reported. The exact reason(s) for the increase is(are) not clear, although the repair and/or construction of wells offered as compensation for damage made to pasture lands by Mongolia's booming mining industry and the introduction of national plans such as the National Livestock Program (approved in 2010 to invest more funds in the livestock sector) may be related. Still, the future of the well network is not yet uncertain and could be a major problem in Mongolia since there is no clear policy for maintaining the well network in rural rangelands.

4.3 Feed supply (vegetation) as a limiting factor for nomadic pastoralism

The discussion above indicates that drinking water supplied from shallow groundwater has been and will likely not be a limiting factor for nomadic pastoralism as long as well networks exist and are maintained. Even if one well dries up, it is possible to move to other areas with a smaller grazing pressure containing a water supply. But does such knowledge imply that increased grazing far above current levels is acceptable and sustainable? In reality, an increase in grazing pressure by 200% is likely not sustainable, not because of the impact on the water balance as has been shown above, but due to a lack of sufficient vegetation to feed the animals. Such a lack of vegetation would result for at least two reasons, as discussed below.

Overgrazing and desertification

A loss of vegetation under heavier grazing will likely result from an increase in animal numbers. For example, Chen et al. (2007) indicated by way of a numerical simulation employing a process-oriented carbon cycle model that both above-ground biomass (AB) and net primary production should keep decreasing and should eventually result in a permanent loss for vegetation if a grazing pressure of S_r > 0.7 SEU ha^{-1} continues over a long time period within the steppe areas of northeastern Mongolia that currently have an annual precipitation rate of 187 mm and a mean S_r of 0.7 SEU ha^{-1}. Thus, the current level of grazing represents the maximum the area can support in a sustainable manner. For the study basins examined, the S_r is 0.56 SEU ha^{-1}, as mentioned above. Since our study areas are located in a drier region than those studied by Chen et al. (2007), the maximum sustainable S_r is expected to be even smaller. Therefore, an increase in the grazing pressure over a much larger level than the current S_r would likely result in the long-term degradation of vegetation.

Water and vegetation

Another potential cause for a loss of vegetation, making it difficult to feed animals, comes from prolonged clear days without significant rainfall events, as was the case in 2011 (Fig. 2). Actually, this type of temporal loss for fresh AB takes place quite frequently because vegetation, particularly herbaceous plants, respond quickly (on daily basis) to water stresses that easily result in the withering of leaves or dormant conditions under continued dry periods. A detailed isotopic study by Satoh et al. (2012) within the W1 watershed indicated such a mechanism. The root systems of herbaceous plants only extend down to 20 cm, so they take water from soils within the upper 20 cm layer. Since this top soil layer dries out very quickly with soil evaporation under rain-free conditions (Fig.2), herbaceous plants are susceptible to water shortages. However, dormant conditions for vegetation without AB can easily be recovered with sufficient rainfall amounts. For example, Satoh et al. (2012) have reported a case in which the re-emergence of AB for herbaceous plants following one month of rainfall absence was observed only two days subsequent to a rainfall event of 6.4 mm and that surface SWC increased from 3 to 5%. Then, following a one week dry period, SWC dropped again to the 3-4% level and a weathering of leaves was observed. However, a total of 6.3 mm rainfall for two continuous days brought SWC up to 9% and the re-growth of AB resulted. Such plant behavior can also be understood from a hydrologic view point using the information listed in Table 4 in which the mean residence time was 1 to 2 months and the mean water storage was only on the order of 10 mm within the upper soil layers from which herbaceous plants take-in water. Thus, it is quite natural that water shortages and recovery processes take place within a relatively short period of time.

All of these arguments and discussions illustrate that much before the shortage of water, feed to livestock itself becomes a problem. Water shortage to vegetation and the resulting loss of pasture will likely emerge as a limiting factor for continuous nomadic pastoralism. Such drought conditions, however, are often limited to certain areas, so nomads can migrate to other areas to seek better pasture conditions as long as the drought is not widespread or the animal number is not too large such that sufficient space is left for migration within the surrounding areas under better conditions. However, for migration to be successful, a well network is necessary so that nomads can move to better areas that contain not only pasture but water supplies from shallow groundwater. Again, the importance of the well network should be stressed.

5. Concluding Remarks

In the present study, the main limiting factors for nomadic pastoralism in Mongolian steppe were investigated by determining water balance and hydrologic processes in two watersheds located in central Mongolia. Since water is often considered to be a limiting factor, a focus was directed towards water availability. In terms of drinking water from shallow groundwater, it was determined that sufficient water exists for consumption by livestock. Such was the case in all seasons throughout a year within the target area, in the past, and, possibly, for the future. Also, based on a comparison with similar studies conducted elsewhere in Mongolia, the results presented are likely applicable to larger areas of Mongolian steppe.

Although shallow groundwater is sufficient, water is more likely to be a limiting factor for the continuation of nomadic pastoralism due to two issues. The first issue is the decline of the well network. As indicated above, the number of wells decreased following introduction of the market-oriented economy in 1990-91, and the maintenance function of wells in Mongolia was weakened. Thus, although shallow groundwater is sufficient, there is a possibility that access to water could be a problem. The second issue to consider is the water-vegetation relationship. For this issue, periods without rainfall result in the temporal loss of aboveground biomass for herbaceous plants that are the main livestock feed source. When a loss of aboveground biomass occurs, a solution that has continued over centuries is simply to migrate to a better area where pasture is abundant due to recent rainfall events. Such is acceptable if the drought condition is not extended to a larger area, and if a well network exists such that nomads can make use of them in order to feed livestock. In this sense, water is an indirect limiting factor for nomadic pastoralism in Mongolian steppe. Interestingly, a study based on a large scale survey of herders by Sternberg (2008) also identified water resources and pasture quality as two of their paramount concerns.

The conclusions presented above were derived based on the first part of our study, namely, water balance estimations over three years in two watersheds in central Mongolia, which can be summarized as follows. The majority (82%) of annual precipitation, P, is lost to evaporation (E), 14% is allotted to soil storage changes (ΔS), and the rest, a small proportion, travels in surface and groundwater runoff. On shorter time scales, the E/P ratio changes greatly and can be >100%. Variation of the E/P ratio is reflected in and absorbed by the change in soil moisture storage, but not that of groundwater in most cases because only the top soil layers are involved in the active hydrologic cycle. Indeed, the residence time of groundwater is much longer (26-55 years) than that of soil water (36-145 days). Thus, recharge to shallow groundwater takes place only for heavy rainfall events within the valley bottom, likely with the occurrence of ephemeral surface runoff. As a result, the soil water-groundwater connection is very weak. Estimated groundwater storage within the shallow aquifer is

S = 420 mm. The amount of the annual withdrawal of shallow groundwater from wells for daily use by livestock is as small as <0.1% of annual precipitation (P), 87.5-95.8 mm.

On a final note, it should be mentioned that, given the remote target area with severe environmental conditions, water balance estimates in the present study are not without uncertainties. Thus, interpretations based on the water balance analysis should also naturally involve uncertainties. Nevertheless, in this study the problem was remedied to some extent by employing multiple approaches in which the water balance components were estimated using two independent methods whenever possible. Although the supplementary information and the results naturally also contain certain errors, all of the available information was consistent and was used to validate the results obtained in the analysis. Another issue that deserves to be mentioned is the representativeness of our detailed study for the two watersheds. As mentioned, the watersheds are located near Mandalgobi and have a mean stocking rate somewhat larger than the average for the larger 'aimag' area, likely due to the closeness of the city. Since a wide variety of conditions exist within the Mongolian steppe region, performing a large-scale survey for validating and supplementing the results presented here is desirable.

Acknowledgement

We thank I. Kaihotsu (Horishima U.) and the Japan Aerospace Exploration Agency for providing wind speed data from their AWS, and G. Davaa and D. Oyunbaatar (Research and Information Institute of Meteorology, Hydrology, and Environment (IMH)) for facilitating local logistics and for arranging the maintenance of stations throughout the Meteorological Centers of IMH at Mandalgobi and Dalanzadgado. We are also grateful to T. Satoh and Y. Morinaga (U. Tsukuba) for participating in the field observations for this study, and to Y. Konagaya (National Museum of Ethnology of Japan), S. Mori (IMG, Inc.), and G.U. Nachinshonhor (Okayama U.) for providing us with references on the history of well construction in Mongolia, and for an anonymous referee and associate editor for providing valuable comments to improve the quality of the manuscript. Finally, we also express appreciation to N. Fujita and N. Yamamura (Research Institute for Humanity and Nature (RIHN)). Without their initiatives and leadership the experiments for this study would not have been possible. Our research was supported and financed, in part, by the RIHN project "Collapse and Restoration of Ecosystem Networks with Human Activity" and by a Grant-In-Aid for Scientific

Research (A, no.24241053) from the Japan Society for the Promotion of Science.

Appendix

Details of the method employed for water balance estimations

Watershed boundaries

Watershed boundaries were determined using a map and GPS surveys. For the surveys, a topographic map of 1:100,000 and a map (Fig.1) created using 90-m digital elevation data from the NASA Shuttle Radar Topographic Mission (SRTM) (CGIAR-CSI, 2008) were employed. The divides determined on these maps by GIS were verified and corrected wherever necessary by in-situ surveys using a GPS receiver. The watershed areas for W1 and W2 were determined to be A = 3.3967 km^2 and A = 11.3567 km^2, respectively.

Soil moisture

For soil water content (SWC) measurements, TDR (time domain reflectmetry) sensors, inter-calibrated prior to installation and installed at the MG1 and MG2 stations, were employed and outputs were corrected based on soil temperature data in order to minimize temperature influences (see Yamanaka et al., 2003). Note that although the SWC measurements were continuous, the SWC measurements obtained during the winter were not as reliable as those obtained during summer because TDR sensors are incapable of detecting accurate soil moisture when they are frozen (e.g., Yoshikawa and Overduin, 2005; Kahimba and Sri Ranjan, 2007; Watanabe and Wake, 2009).

Assumptions in water balance estimations

As mentioned in the methods section, for application of the water balance equation (1) we assumed that the shallow aquifer is underlain by an impermeable layer and that no leakage occurs. Based on the steady groundwater level, the assumption was judged to be acceptable. However, there is a possibility that the excessive withdraw of deep groundwater may induce the leakage of shallow groundwater down to deeper aquifers (e.g., Rathod and Rushton, 1991; Yamanaka et al., 2011). Such could be the case in Mandalgobi since municipal water for the city comes from three wells located some 30 km from the city and three wells within the city (Tuinhof and Nemer, 2009) with depths of 160 m (Tsogtbaatar and Janchivdorj, 2009). Perhaps the amount of extracted water (10^3 m^3/day during winter and 7.5×10^2 m^3/day during summer (Tuinhof and Nemer, 2009)) is not large enough to cause large-scale induced seepage.

Precipitation

As mentioned, P was determined from AWS measurements. However, the tipping bucket rain gauge at MG1 was not capable of measuring snowfall during the winter. Thus, to supplement AWS measurements, daily precipitation data measured manually at the IMH station were also used for considering the water balance of each watershed.

Evaporation

Evaporation, E, was estimated from the following energy balance equation:

$$R_n - G = H + LE \tag{A.1}$$

where R_n is net radiation, G is the soil heat flux, LE (= $L_e \times E$) is the latent heat flux, and L_e is the latent heat for vaporization. H is the sensible heat fluxand was estimated by applying the following bulk equation for sensible heat flux:

$$H = \frac{k u_* \rho c_p (T_s - T)}{\ln\left(\frac{z - d_0}{z_{0h}}\right) - \Psi_h\left(\frac{z - d_0}{L}\right)} \tag{A.2}$$

where k is Karman's constant; u_* is the friction velocity; ρ is the air density; c_p is the specific heat of air at constant pressure; T_s and T are the surface and air temperature, respectively; z is the height of the T measurement; and d_0 is the displacement height, assumed to be 2/3 of the mean vegetation height (= 0.145 m for *Caragana*) for MG1 and zero for MG2. The symbol z_{0h} represents scalar roughness for sensible heat, Ψ_h is the stability correction function for sensible heat, and L is the Obukhov length. The frictional velocity, u_*, was derived by the bulk equation for momentum flux, as follows:

$$u_* = \frac{ku}{\ln\left(\frac{z - d_0}{z_0}\right) - \Psi_m\left(\frac{z - d_0}{L}\right)} \tag{A.3}$$

where u is the wind speed, z_0 is the roughness length, and Ψ_m is the stability correction function for momentum.

Surface parameters for z_0 and z_{0h} were first determined from (A.2)-(A.3) using fluxes for H, LE, and u_* obtained from an eddy correlation approach. More specifically, the method of Toda and Sugita (2003) was applied to the data, in which the left hand side (LHS) (i.e., the measured fluxes) and the right hand side (RHS) of (A.2)-(A.3) (i.e., the estimated fluxes) for a given value of z_0 and z_{0h} were compared. The process was repeated by changing z_0 and z_{0h} over a small interval, and parameters that yielded the smallest rms error were selected. For the present analysis, rms error, the statistics of R^2, the ratio of the means for LHS and RHS, and the regression constants of a and b in $\hat{y} = ax + b$ (where $\hat{y}$ is the measured flux and x is the estimated flux) were considered.

To apply this approach, intensive observations were performed during the summer of 2009; and an eddy correlation system with a sonic anemometer (Gill Instruments, Ltd., R3-50) and an open-path gas analyzer (LI-COR Inc., LI-7500) was added to the AWS system from 23-28 July at MG1 and from 28-30 July at MG2 in order to determine the surface roughness parameters of the area so that the bulk approach for estimating surface fluxes could be applied (see below).

Based on observation results, $z_0 = 1.73\times10^{-3}$ m, and $z_{0h} = 1.71\times10^{-5}$ m were determined for the area surrounding the MG1 station. The values agree with the theoretical estimates for bluff-rough surfaces (e.g., Fig. 4.24 of Brutsaert (1982)), more so than those for a surface covered with permeable roughnesses. The finding is not surprising since the average fractional coverage of *Caragana* was only 3-4% (Satoh et al., 2012) and the surface is closer to that of bare soil than a vegetated surface. The same values of z_0 and z_{0h} were used for the entire year and for MG2 following verification that the values produced fluxes that agreed, on average, with measured fluxes during the intensive observation period.

The soil heat flux, G, was estimated as a fraction of R_n. Constants (0.68 for daytime and 1.1 for nighttime) were determined based on measured values for H and LE obtained from the eddy correlation approach and R_n from net radiometer data obtained during intensive measurements. G values were estimated as the residual term of the energy balance equation from $G = R_n - LE - H$. Although it is possible to use the G obtained from a soil heat flux plate, this approach was rejected based on a preliminary analysis in which closure of the energy balance (A.1) in the form of $y = ax + b$ and the ratio of the means $\bar{y}/\bar{x}$ in which $x = R_n - G$ and $y = H + LE$ (e.g., Wilson et al., 2002) was evaluated. The result was a = 0.43, b = -50, and $\bar{y}/\bar{x}$ = 0.42 and clearly indicated the presence of a larger imbalance in the energy balance than the values reported in the literature. Following a careful check of each term, we suspected that the soil heat flux, G, measured by the soil heat flux plate was not representative of the area and resulted in underestimations that caused an imbalance in the energy balance. Due to this finding, we decided not to use the measured G.

With measurements and parameters, it is possible to continuously estimate fluxes from 30-min averaged AWS measurements. However, for the period from April through July of 2009 we encountered a problem with the wind speed sensor at MG1. Due to the malfunction, wind speed values during this period were estimated from hourly wind speed measurements obtained from a nearby AWS operated by the Univ. Hiroshima. For estimations, a conversion equation was derived based on measurements obtained when both AWSs were operating simultaneously.

Since (A.1)-(A.3) are implicit, an iteration procedure was applied in order to derive the fluxes of H, u_*, and LE. Thus, $L = \infty$ was assumed for deriving the first estimate for H and u_*, and LE from (A.1). The procedure produced the first estimates for L and allowed determinations of Ψ_m, Ψ_h, and a second round of estimates for H, u_*, and LE, etc. The procedure was repeated until fluxes sufficiently converged. The derived fluxes for MG1 and MG2 were found to be quite similar. Therefore, we decided to use the fluxes obtained for the MG1 stations for additional calculations for both watershedsin what follows. This has an advantage for having evaporation over longer time periods.

Groundwater level

The water table level of the three wells, GW1, GW2, and GW4, was manually measured and recorded daily by local herders. Manual recording was chosen due to the difficulty in recovering automatic sensors in remote areas, especially after a full year. Manual water level data were quality checked in order to screen (and correct when possible) suspicious data. An example of a typical correction occurred due to shifts in water level prior to and following the winter season because during the winter well water was frozen and no measurements were performed. We also followed-up on manual reports when we suspected that the reference for ground level was somehow moved and caused a shift in the water level. An additional example of data correction occurred due to unrealistic fluctuations for water levels reported during the first year. We corrected first year data during the second year after issuing instructions for more careful measurements.

Groundwater discharge

For the W1 watershed, groundwater discharge can be assumed to be zero because GW1 is located at the center of a local depression near the watershed boundary. From this depression there is no apparent outlet towards the downstream direction.

For the W2 watershed, estimates of G_{out} were obtained by applying Darcy's law, as follows:

$$G_{out} = q_x \frac{A_c}{A} = -k_h \frac{dh}{dl}\frac{A_c}{A} \cong -k_h \frac{h_2 - h_1}{l_{2-1}}\frac{A_c}{A} \qquad (A.4)$$

where A_c is the cross-sectional area of groundwater discharge from a watershed of area A, q_x is the specific groundwater flux, k_h is the hydraulic conductivity, and dh/dl is the hydraulic gradient. The value of dh/dl was estimated from the difference for the hydraulic head of observational wells GW2 and GW4 located along the direction of the main discharge direction as implied by surface topography over a horizontal distance of l_{2-1}.

The value of k_h was determined by applying the Theis well recovery test (e.g., Kruseman and de Ridder, 1990) to wells GW1 and GW4. All water within the wells was first emptied, then water recovery was recorded in order to estimate transmissivity, then k_h for an aquifer with a known thickness in and around the W1 and W2 watersheds was determined. From tests performed from 25-31 July 2009 at GW1 and 28 July at GW4, values of $k_h = 2.1\times10^{-8}$ m s^{-1} for GW1 and $k_h = 2.7\times10^{-7}$ m s^{-1} for GW4 were obtained. The obtained k_h values are in the general range of those expected for an aquifer with unconsolidated materials of silty sand or silt (e.g., Freeze and Cherry, 1979). For estimations of G_{out} for the W2 watershed, the k_h value was determined based on the adopted GW4 well recovery test. Cross-sectional area was determined from a GPS survey and from information on the thickness of the aquifer as well as the mean water table depth.

Surface runoff

Surface runoff, R_{out}, can be assumed to be zero for the W1 watershed for the same reason that $G_{out} = 0$ and because no gully formation was found within the local topography of the watershed outlet. In contrast, there was a clear indication of surface runoff within the topography of the W2 outlet. Therefore, R_{out} was estimated based on the following consideration. Interviews with the local population allowed us to obtain crude estimations for the frequency and magnitude of surface runoff occurrences. From interviews we learned that W2 surface runoff at the outlet of the catchment (i.e., at around GW2) had been observed only once every 2-3 years and that when it occurred it had a depth of several centimeters and lasted, at most, several hours. We then compared the report of locals with AWS rainfall data. Over approximately three summer seasons the largest recorded rainfall events had an intensity of 4.8 and 6.3 mm within a ten minute period. A rainfall intensity of >2.5 mm 10 min^{-1} occurred five times during the same interval. Thus, a value of ~4-5 mm 10 min^{-1} could likely be adopted as a threshold value above which overland flows result. Interestingly, Onda et al. (2007) reported, based on observations for two years in two hilly watersheds located in northeastern Mongolia, that a threshold value of 4 mm 10 min^{-1} initiated surface runoff. Note that rainfall events with an intensity of 4.8 and 6.3 mm 10 min^{-1} took place on days with daily rainfall totals of 10.0 and 18.5 mm, respectively, that have a return period of 0.9 and 1.7 years, respectively, as determined from annual maximum daily rainfall amounts from the IMH station over 65 years fitted using the Weibull plotting position formula (e.g., Brutsaert, 2005). Similarly, the 2-year and 3-year rainfall values were, respectively, 21 and 27 mm day^{-1}.

The maximum depth (within the stream cross-section) of surface runoff and the duration of runoff were assumed to have values of 2 cm, 5 cm, and 10 cm, and 1 to 3 hours to examine their sensitivity to water balance. However, the use of different depths and duration did not produce significantly different results.

The mean velocity for surface runoff was estimated by applying the Gauckler-Manning equation (e.g., Brutsaert, 2005). Based on visual observations, the roughness parameter n = 0.022 was used for a straight channel with a sandy river bed. The mean channel gradient (0.016) and the cross sectional shape of 11 points along the channel from the beginning of the gully toward the downstream over a distance of 0.925 km were determined from the in-situ survey. The mean hydraulic radius and the cross-sectional area of flow were then determined from the wetted areas of the 11 cross sections for the assumed depths. The values are 6.3 cm and 0.36 m^2, respectively, for the assumed 10-cm depth.

Water withdrawal from wells by local herdsmen

Water withdrawal by local herdsmen was estimated for the current animal numbers (Table 2) obtained from annual official statistics in 2007 (as compiled by the National Statistical Office), from information provided by local herdsmen in 2009, and from unit water consumption values for which several sources were available. For example, Baranchuluun et al. (2004) listed water use values for different seasons and/or for the different life stages of livestock. The Minister of Nature and Environment (1995) also provides standard water consumption values for the summer and winter seasons. Although there are some differences in the listed values in these references, for our purpose of estimating annual U values, the mean annual unit consumption of water for each livestock type was derived and used for the analysis (Table 2).

Groundwater Storage

Groundwater storage, *S*, was estimated based on the thickness and effective porosity of the aquifer. The effective porosity value was assumed to be the same as the mean specific yield (0.3) of 163 wells (depth <40 m) in Saintsagaan soum. The result appears to be on the higher side for an aquifer of sandy to silty materials. The mean thickness of the aquifer was first determined by assuming that the aquifer thickness for the location of the wells was 5.1 m, the average aquifer thickness for 15 wells registered in and around the W1 and W2 watersheds in the Well Database. We also assumed that thickness decreases linearly along the topographic slope from the valley bottom toward the watershed boundary (the topographic divide) where the thickness was assumed as zero. More specifically, aquifer depth was first determined manually at approximately 20 points within the W1 and W2 watersheds along the slope. The Kriging interpolation was then applied together with zero thickness information on the divide in order to determine the distribution for thickness within the watershed. The mean aquifer depth of each watershed was determined to be 1.4 m for W1 and 1.3 m for W2, which, in turn, produced the amount of storage for each watershed.

Storage change

In theory, it should be possible to derive storage change from the measured water levels of the observation wells and soil moisture profile measurements by assuming that they represent the entire basin. However, as mentioned above, well water level data as measured manually by local herders were judged not to be as accurate for determining water level changes as had been hoped. However, what these data did indicate was that water levels were very stable without a clear seasonal trend and that response to rainfall events and daily water withdrawals, in most cases, were not pronounced. Thus, assuming that the storage change in groundwater is usually negligible is a reasonable assumption. For the storage change within soil water, SWC values measured at the MG1 and MG2 stations were used for estimating ΔS from the surface down to -1.3 m. Since the average water table depth in this area is 2.1 m (as obtained from the Well Database of Mongolia) and since the SWC value did not change much at -1.1 m (see the Results section), ΔS determined this manner should provide acceptable estimates for storage changes within the soil layer.

Residence time of groundwater

If one can assume piston flow for groundwater, the mean residence time, $\bar{t}$, of shallow groundwater and the tritium concentration, C, are related by the following equation (e.g.,Ozyurt and Bayari, 2005):

$$C = C_0 \exp\left(-\bar{t}/T\right) \qquad \text{(A.5).}$$

where C_0 is the tritium concentration at the time of groundwater recharge and T is the average lifetime of tritium (≈17.8 years). The value of C_0 can be estimated using the past record of tritium concentration data for rainfall in the region and the radioactive decay of tritium. The former time series data were constructed in Higuchi (2005) using data obtained in Ulaanbaatar in Mongolia (1990-2000), Havarovsk in Russia (1970-1983), and Ottawa in Canada (1970). Values for C_0 were visually estimated from Fig. A1 as 80-200 TU for GW1 and 50-100 TU for GW2 and GW4, from which the mean residence time, $\bar{t}$, of shallow groundwater was estimated (Fig. A1 and Table 5).

Reference

Anonymous, 1982. *The Secret History of the Mongols Vol. 1* (translated by Cleaves, F. W.), Harvard Univ. Press, Cambridge, 277p.

Arora, V.K. 2002. The use of the aridity index to assess climate change effect on annual runoff. *J. Hydrol.*, 265:164-177

Baranchuluun, Sh., Chanlmanii, L., Lugarmaa, M., 2004. *Water Supply in Pasturelands* (Баранчулуун, Ш.,Чандмань, Д. and ДугармааМ. (2004), Бэлчээр усжуулалт, , Дархан-Сэлэнгэ ХХК, 183р.) in Mongol.

Brutsaert, W., 1982. *Evaporation into the Atmosphere: Theory, History, and Application*, D. Reidel Pub. Co., 299p.

Brutsaert, W., 2005, *Hydrology: An Introduction*, Cambridge Univ. Press, 605p.

Brutsaert, W., Sugita, M. 2008. Is Mongolia's groundwater increasing or decreasing? The case of the Kherlen River vasin. *Hydrol. Sci. J.*, 53, 1221-1229. doi: 10.1623/hysj.53.6.1221

Byambakhuu, I., Sugita, M., Matsushima, D., 2010. Spectral unmixing model to assess land cover fractions in Mongolian steppe regions. *Remote Sens. Environ.*, 114, 2361-2372 doi:10.1016/j.rse.2010.05.013.

Chen, Y., Lee, G. Lee P., Oikawa, T., 2007. Model analysis of grazing effect on above-ground biomass and above-ground net primary production of a Mongolian grassland ecosystem. *J. Hydrol.*, 333, 155-164, doi:10.1016/j.jhydrol.2006.07.019.

CGIAR-CSI, 2008. *SRTM 90 m Digital Elevation Data*, http://srtm.csi.cgiar.org/Index.asp. Accessed 12 April 2009

Chuluun T., Ojima D., 2002. Sustainability of pastoral systems in Mongolia, in *Fundamental Issues Affecting Sustainability of the Mongolian Steppe*, ed. by Chuluun, T. and Ojima, D., Interpress, Ulaanbaatar, 52-57.

Cribb, R., 1991. *Nomads in Archaeology*, Cambridge Univ. Press, 253p.

Dorzhgotov, D., 2003. *Soils of Mongolia*. Admon, Ulaanbaatar, 287p. (in Mongol).

Freeze, R.A., Cherry, J.A., 1979. *Groundwater*. Prentice-Hall, 604p.

Healy, R.W. 2010. *Estimating Groundwater Recharge*. Cambridge Univ. Press. 256p.

Higuchi S. 2005. *Groundwater Flow System Study Revealed by Environmental Tritium and 3-D Groundwater Simulation in Kherlen River Basin, Mongolia*, MS Thesis, Graduate school of Science and Technology, Kumamoto University, 89p. (in Japanese).

Jacoby, G. C., D'Arrigo, R. D., Pederson, N., Buckley, B. M., Dugarjav, Ch., Mijiddorj, R., 1999. Temperature and precipitation in Mongolia based on dendroclimatic investigations, *IAWA J,* 20, 339-350.

Kahimba, F.C., Sri Ranjan, R., 2007. Soil temperature correction of field TDR readings obtained under near freezing conditions, *Can. Biosys. Eng*., 49, 1.19-1.26.

Kruseman, G.P., de Ridder, N.A., 1990. Analysis and Evaluation of Pumping Test Data, Revised Edition, International Institute for Land Reclamation and Improvement, Publ. 46, 377p.

Matsubara, M., 2005. Introduction: toward the compilation of Eurasian history, in Matsubara, M., Konagaya, Y., and Yang, H. ed., 2005. *Message from Eurasian Grasslands*, Heibonsha, Ltd., Tokyo, 11-29.

Mineral Resources Authority of Mongolia and Mongolian Academy of Sciences, 1998. G*eological Map of Mongolia, Scale 1:1,000,000*, Ulaanbaatar, available on line at Mongolian Mining Registry and Cadastre System (URL: http://www3.gaf.de//Mongolia _mining/pages/pub/downloads.html)

Mineral Resources Authority of Mongolia, 2001. *Distribution of Deposits and Occurrences in Mongolia (at the scale of 1:1,000,000)*, Ulaanbaatar, available on line at Mongolian Mining Registry and Cadastre System (URL: http://www3.gaf.de//mongolia_mining/pages/pub/downloads.html)

Minister of Nature and Environment, 1995. *Water Use Statistics by Temporal Regulation (Norm or Standard)*, Ministry of Nature and Environment, Order No. 153, 153p.

Miyawaki, J., 2002. *History of Mongolia*, Tosui Shobo, Tokyo,295p. (in Japanese).

National Statistical Office of Mongolia 1999, *Yearbook 1999*, National Statistical

Office of Mongolia, available online at http://www.nso.mn

National Statistical Office of Mongolia, 2005. *Yearbook 2005*, National Statistical Office of Mongolia, available online at http://www.nso.mn

National Statistical Office of Mongolia, 2007. *Yearbook 2007*, National Statistical Office of Mongolia, available online at http://www.nso.mn

National Statistical Office of Mongolia, 2009. *Yearbook 2009*, National Statistical Office of Mongolia, available online at http://www.nso.mn

Onda, Y., Kato, H., Tanaka, Y., Tujimura, M., Dvaa, G., Oyunbaatar, D., 2007. Analysis of runoff generation and soil erosion processes by using environmental radionuclides in semiarid areas of Mongolia, *J. Hydrol.*, 333, 124-132, doi:10.1016/j.jhydrol.2006.07.030.

Ozyurt, N.N., Bayari, C.S., 2005. LUMPED Unsteady: a Visual Basic code of unsteady-state lumped-parameter models for mean residence time analyses of groundwater systems, *Computers Geosci.*, 31, 329-341.

Pederson, N., Jacoby, G. C., D'Arrigo, R. D., Cook, E. R., Buckley, B. M., Dugarjav,C., Mijidorj, R., 2001. Hydrometeorological reconstructions for northeastern Mongolia derived from tree rings: 1651-1995, *J. Climate,* 14, 872-881.

Pratt M., Rasmussen, G.A., 2001. *Determining Your Stocking Rate*. Utah State University Extension Electric Publishing, NR/RM/04, 7p.

Rathod, K. S., Rushton, K.R., 1991. Interpretation of pumping from two-zone Layered aquifers using a numerical model, *Groundwater*, 29, 499-509. doi: 10.1111/j.1745-6584.1991.tb00541.x

Sasaki T., Okayasu, T. Jamsran, U., Takeuchi, K., 2008. Threshold changes in vegetation along a grazing gradient in Mongolian rangelands, J. Ecol., 96, 145-154, doi: 10.1111/j.1365-2745.2007.01315.x.

Sato, T., Kimura, F., Kito, A., 2007. Projection of global warming onto regional precipitation over Mongolia using a regional climate model. *J. Hydrol.*, 333, 144-154, doi: 10.1016/j.jhydrol.2006.07.023

Satoh, T., Sugita, M., Yamanaka, T., Tsujimura, M., Ishii, R., 2012. Water dynamics within a soil-vegetation-atmosphere system in a steppe region covered by shrubs and herbaceous plants, in: Yamamura N, Fujita N, Maekawa A (eds), *The Mongolian Network: Environmental Issues under Climate and Social Changes. Ecological Research Monographs,* 43-63, Springer, Tokyo.

Scanlon, B.R., Keese, E., Flint, L., Flint, L.E., Gaye, C.B., Edmunds, W. M., Simmers, I., 2006. Global synthesis of groundwater recharge in semiarid and arid regions. Hydrol. Processes, 20, 3335–3370, doi: 10.1002/hyp.6335.

Sharkhuu, Ts., 1975. *The Situation of Land Communication Before the Revolution in Mongolia*. (Хувьсгалын өмнөх Монгол дахь газрын харилцаа, Улсын хэвлэлийн газар), State Publication House, 252p (in Mongol)

Shiirevdamba, Ts., 1998. *Biological Diversity in Mongolia, First National Report*, Ministry of Nature and Environment, 106p. Ulaanbaatar, Mongolia: Admon Printing House.

Sternberg, T., 2008. Environmetal challenges in Mongolia's dryland pastoral landscape. *J. Arid Environ.*, 72, 1294-1304. doi:10.1016/j.jaridenv.2007.12.016

Sugita, M., Asanuma, J., Mariko, S., Tsujimura, M., Kimura, F., Lu, M., Azzaya D. and Adyasuren Ts., 2007. An Overview of the Rangelands Atmosphere Hydrosphere Biosphere Interaction Study Experiment in Northeastern Asia (RAISE). *J. Hydrol.*, doi:10.1016/j.jhydrol.2006.07.032.

Toda, M. Sugita, M., 2003. Single level turbulence measurements to determine roughness parameters of complex terrain, *J. Geophys.Res.*, 108, D12, 4363, doi: 10.1029/2002JD002573

Tsogtbaatar, J. Janchivdorj, L., 2009. *The Final Report of the Project on "Determination of Heavy Metal in Surface Water and Groundwater of Mongolia"*, Institute of Geoecology, Mongolian Academy of Sciences, 60p.

Tsujimura, M., 2007. How much groundwater is available in a well, *Bull. Terr.Environ. Res. Ctr., Univ. Tsukuba*, No.8 Supplement, 82-83.

Tuinhof, A., Nemer, B., 2009, *Groundwater Assessment in the Gobi Region: Final Draft Report 25/01/2009.* Acacia Water, 56p. and Annex 1-7.

Watanabe, K. Wake, T., 2009. Measurement of unfrozen water content and relative permittivity of frozen unsaturated soil using NMR and TDR, *Cold Regions Sci. Technol.*, 59, 34-41.

Wilson, K.B., Goldstein, A.H., Falge, E., Aubinet, M., Baldocchi, D., Berbigier, P., Bernhofer, Ch., Ceulemans, R., Dolman, H., Field, C., Grelle, A., Law, B., Meyers, T., Moncrieff, J., Monson, R., Oechel, W., Tenhunen, J., Valentini, R., Verma, S., 2002. Energy balance closure at FLUXNET sites. *Agricul. For. Meteorol.*, 113, 223–243.

WMO (World Meteorological Organization), 2012. *Standardized Precipitation Index; User Guide*, WMO-No.1090, World Meteorological Organization, Geneva.

Yamanaka, T, Kaihotsu, I, Oyunbaatar, D., 2003. Temperature effect on soil water content measured by time domain reflectometry and its correction using in situ data. *Bull. Terr. Environ. Res. Ctr., Univ. Tsukuba*, 16, 246-254 (in Japanese).

Yamanaka, T., Kaihotsu, I., Oyunbaatar,D., Ganbold, T., 2007. Summertime soil hydrological cycle and surface energy balance on the Mongolian steppe, *J. Arid Environ.*, 69, 65-79.

Yamanaka, T., Shimada, J., Tsujimura, M., Lorphensri, O., Mikita, M., Hagihara, A, Onodera, S., 2011. Tracing a confined groundwater flow system under the pressure of excessive groundwater use in the Lower Central Plain, Thailand. *Hydrol. Processes*, 25, 2654-2664. doi:10.1002/hyp.8007.

Yoshikawa, K., Overduin, P.P., 2005. Comparing unfrozen water content measurements of frozen soil using recently developed commercial sensors. *Cold Regions Sci. Technol.*, 42, 250-256.

Table 1 A list of variables measured at the two stations.

Variable	Station	Height (m)	Sensor	Remarks
Wind speed	MG1	3.0	R.M. Young, 3101	No measurements in April-July of 2009 (see Appendix)
Temperature and relative humidity	MG1	2.65	Campbell Scientific, 41003-5 and Vaisala, HMP45D	The prove is housed in a radiation shield that has natural ventilation
Radiation	MG1	2.5	Hukseflux Thermal Sensors, NR01	Four radiation components (up and down, short and log-wave)
	MG2	2.5	Hukseflux Thermal Sensors, RA01	Two components (upward only)
Soil heat flux	MG1	-0.01	Hukseflux Thermal Sensors, HFP01	
Rainfall	MG1	0.6	Texas Electronics, TR-525M	Tipping bucket rain gauge
Soil water content (SWC)	MG1	-0.05, -0.1, -0.2, -0.3, -0.7, -1.1 (under a *Caragana* mound) and -0.05, -0.1, -0.2, -0.3, -0.5, -0.7, -1.1 (only from 2010/5, under *Allium* surface)	Campbell Scientific, CS-616	Time domain reflectometry (TDR) sensor
	MG2	-0.1, -0.2, -0.3, -0.5, -0.7		
Soil temperature	MG1	-0.1, -0.3, -0.7		Platinum resistance thermometer
	MG2	-0.1, -0.2, -0.3, -0.5, -0.7		

Table 2 Number of livestock in 2008 and assumed unit water consumption.

	W1 (heads)	W2 (heads)	Saintsagaan Soum (heads)	Unit water consumption (L $head^{-1}$ day^{-1})
Sheep	73	200-230	94,024	4
Goats	151	200-300	107,575	4
Horses	0	14	8,732	29
Cattle	0	0	2,794	31
Camels	0	16-20	761	60

Table 3 Annual water balance for the three selected years.

Period	I		II		III		Average of periods I, II and III	
	2008/7/8-2009/7/7		2009/5/4-2010/5/3		2010/7/3-2011/7/2			
	W1	W2	W1	W2	W1	W2	W1	W2
P	95.5		87.5		95.8		92.9	
E	65.1 (68)		83.7 (96)		80.7 (84)		76.5 (82)	
G_{out}	0	4×10^{-5} (10^{-5})	0	4×10^{-5} (10^{-5})	0	4×10^{-5} (10^{-5})	0	4×10^{-5} (10^{-5})
R_{out}	0	0-0.3 (0-0.3)	0	0	0	0-0.3 (0-0.3)	0	0-0.2 (0-0.2)
U	$9.6\ 10^{-2}$ (10^{-1})		$9.6\ 10^{-2}$ (10^{-1})		$9.6\ 10^{-2}$ (10^{-1})		$9.6\ 10^{-2}$ (10^{-1})	
ΔS	28.8 (30)		28.0 (32)		-18.7 (-20)		12.7 (14)	
ε	1.2-1.5		-24.3		33.4-33.7		3.3 ~ 3.6	
PE	1160.0		1038.3		1029.5		1075.9	

The unit is mm a^{-1} and the percentage of each component to total *P* is given in parentheses. The error, ε, *was* determined by $\varepsilon = P - E - G_{out} - R_{out} - U - \Delta S$. The range of R_{out} for W2 is given for the assumed case of (1) with a depth, *h* = 0.02 cm, a duration, *t* = 1 h; (2) with a *h* = 0.10 cm and a *t* = 3 h; and both with a threshold rainfall intensity of P_i = 4 mm 10 min^{-1}.

Table 4 Values of $P-\Delta S$, mean storage, $\overline{S}$, and mean residence time, $\bar{t}$, estimated for the various soil layers.

MG1				MG2			
Depth (m)	$P-\Delta S$ (mm)	$\overline{S}$ (mm)	$\bar{t}$ (days)	depth (m)	$P-\Delta S$, (mm)	$\overline{S}$ (mm)	$\bar{t}$ (days)
0-0.075	56.7 (70.4)	4.7	25				
0-0.15	62.6 (77.7)	9.4	49	0-0.15	58.5 (87.6)	6.9	36
0-0.25	68.5 (85.1)	14.8	78	0-0.25	60.9 (91.3)	10	52
0-0.5	80.3 (99.8)	32	167	0-0.4	62.8 (94.5)	15.9	83
0-0.9	80.1 (99.5)	58.8	308	0-0.6	64.0 (95.8)	20.2	106
0-1.3	80.5 (100)	87.2	457	0-0.8	66.8 (100)	27.8	145

The number inside parentheses indicates the percentage of $P-\Delta S$ for each depth range to $P-\Delta S$ for the largest depth range for each station. $P-\Delta S$ was determined for 2009/5/4-2010/3/8 for the MG1 station, and for 2009/5/4-2010/3/31 for the MG2 station in order to avoid periods with possibly large errors for $P-\Delta S$. The mean residence time was estimated by $\bar{t}=\overline{S}/\overline{P}$ where $\overline{P}$ is mean daily precipitation.

Table 5 Results based on the tritium analysis of well water.

	C (TU)	C_0 (TU)	$\bar{t}$ (years)	$\overline{q_z}$ (mm a^{-1})
GW1	17.19±2.32	80-200	28-44	9.5-15
GW2	3.70±1.50	50-100	47-59	7.1-8.9
GW4	7.76±1.34	50-100	34-46	9.1-12

C is the tritium concentration of sampled well water, C_0 is the estimated tritium concentration of groundwater at the time of recharge, $\bar{t}$ is the mean residence time of shallow groundwater derived from (A.5), and $\overline{q_z}$ is the mean recharge rate to shallow groundwater estimated by $\overline{q_z} = \overline{S} / \bar{t}$.

Figure caption

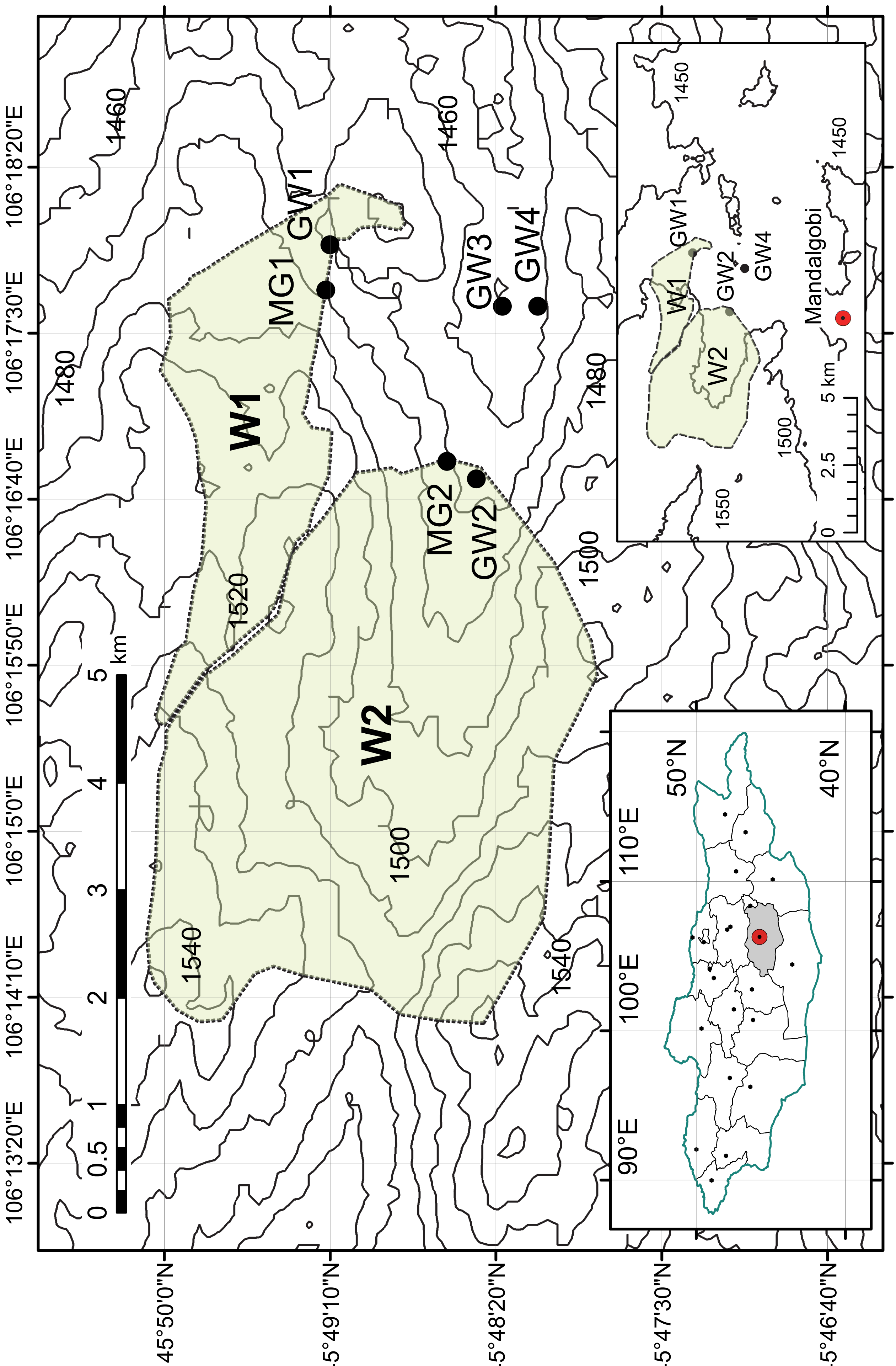
Figure 1 The topography of the study area. Closed circles indicate the locations of automatic weather stations (AWS) MG1 and MG2 and wells GW1-GW4. Watersheds boundaries are shown with dotted lines, while contours are given by continuous lines at 10 m intervals. The smaller watershed located to the east is W1 and the larger watershed to the west is W2. Also shown as an inset is a map of Mongolia with aimag (prefecture) boundaries and capital cities, and those of the surrounding area of the two watersheds. The study area is located near Mandalgobi in Dundgobi Aimag (shown in the map as a red circle with a center dot and as the shaded area).

Figure 2 The time change for groundwater level (GWL, m asl), soil temperature (T_{soil}) at a depth of 10 cm, soil water content (SWC), and the accumulated values of *E, P*, and $P-\Delta S$ in which ΔS in this case was determined from SWC data obtained from the surface down to -1.3 m. All values were obtained from the MG1 AWS station, with the exception of GWL. Some of the MG1 AWS station data were not available for the period of 2010/5/12-7/2, due to a technical problem, and, thus, the accumulated lines of *E, P*, and $\sum P-\Delta S$ were reset to zero. The dotted line for SWC data is for measurements under an *Allium* surface. The depth of the sensor is indicated by the color of each line.

Figure 3 Changes in livestock number in Saintsagaan soum and the number of wells within Mongolian pastureland based on the National Statistical Office of Mongolia (1999, 2005, 2007, and 2009).

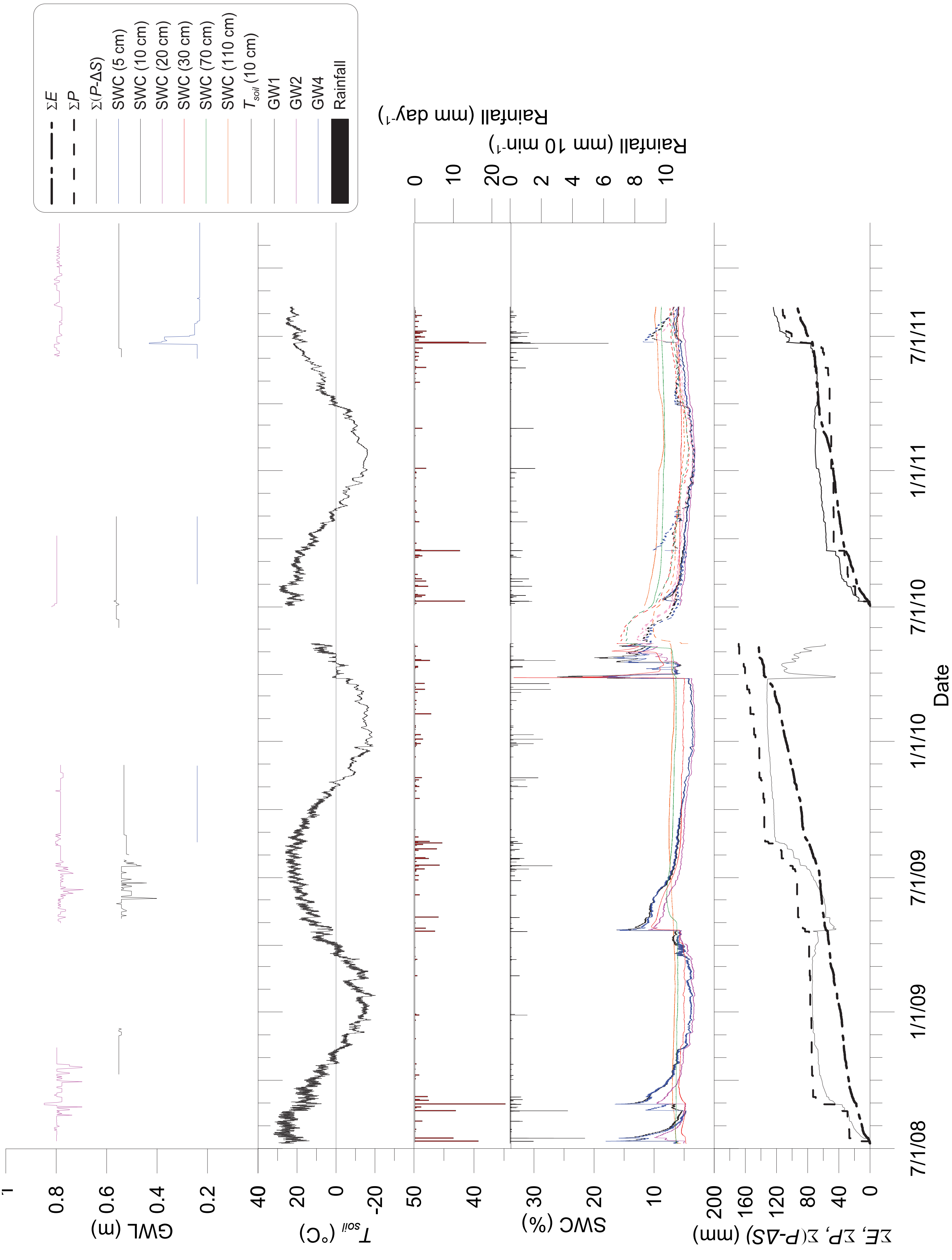
Figure A1 The time variation for the precipitation tritium concentration (open circles) and groundwater (closed triangles). Dotted lines represent the decay function $N = N_0 \exp(-\lambda t)$ through the tritium concentration for water samples from GW1, GW2, and GW4 during 2011. t is the elapsed time and N is the number of nuclides in T.U., λ is the decay constant, and N_0 is N at an initial time of $t = 0$.

Fig. 1



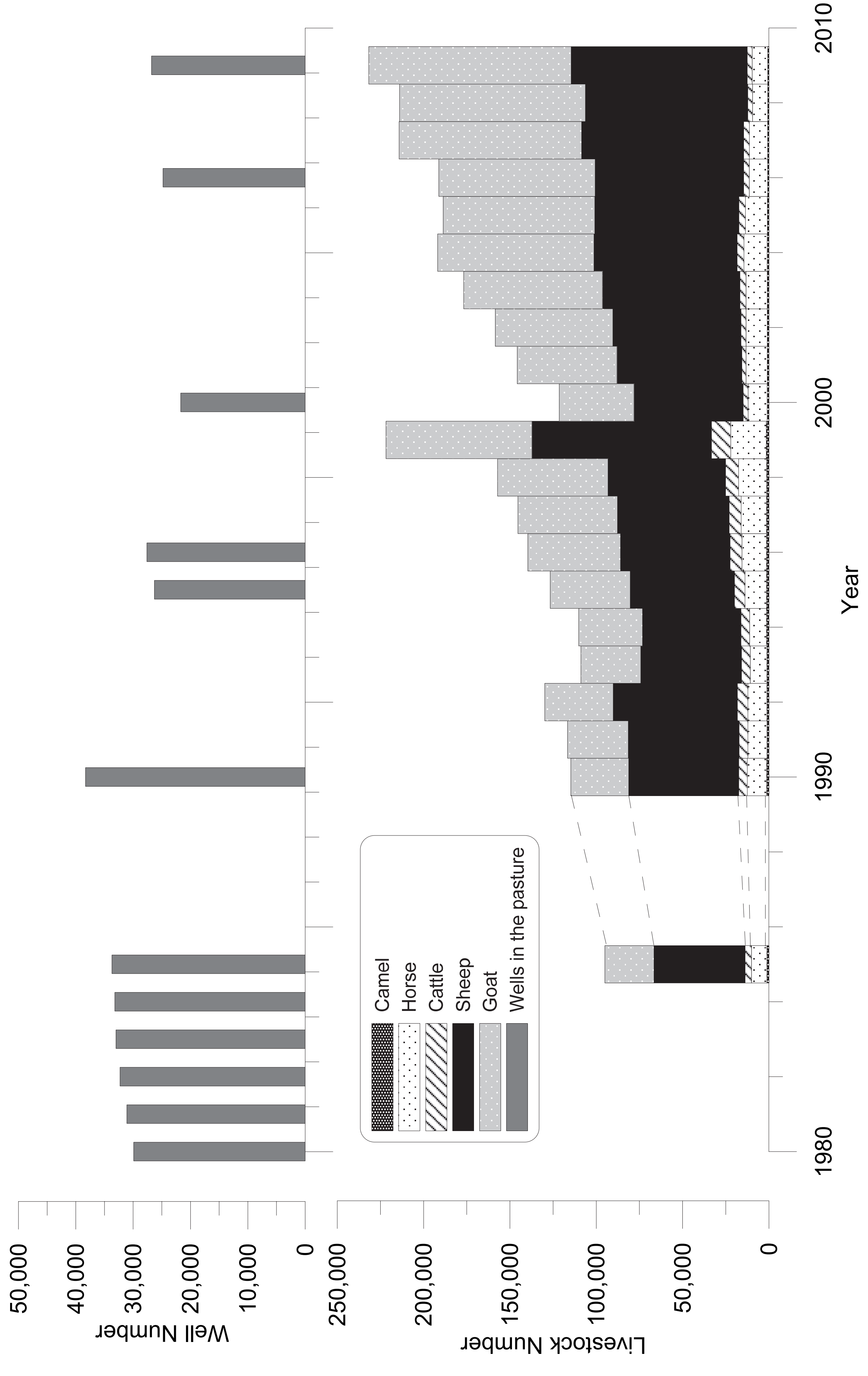
W1
W2
MG1
MG2
GW1
GW2
GW3
GW4
Mandalgobi
1460
1480
1500
1520
1540
1450
1550
0 0.5 1 2 3 4 5 km
0 2.5 5 km
106°13'20"E
106°14'10"E
106°15'0"E
106°15'50"E
106°16'40"E
106°17'30"E
106°18'20"E
45°50'0"N
45°49'10"N
45°48'20"N
45°47'30"N
45°46'40"N
90°E
100°E
110°E
40°N
50°N

Fig. 2



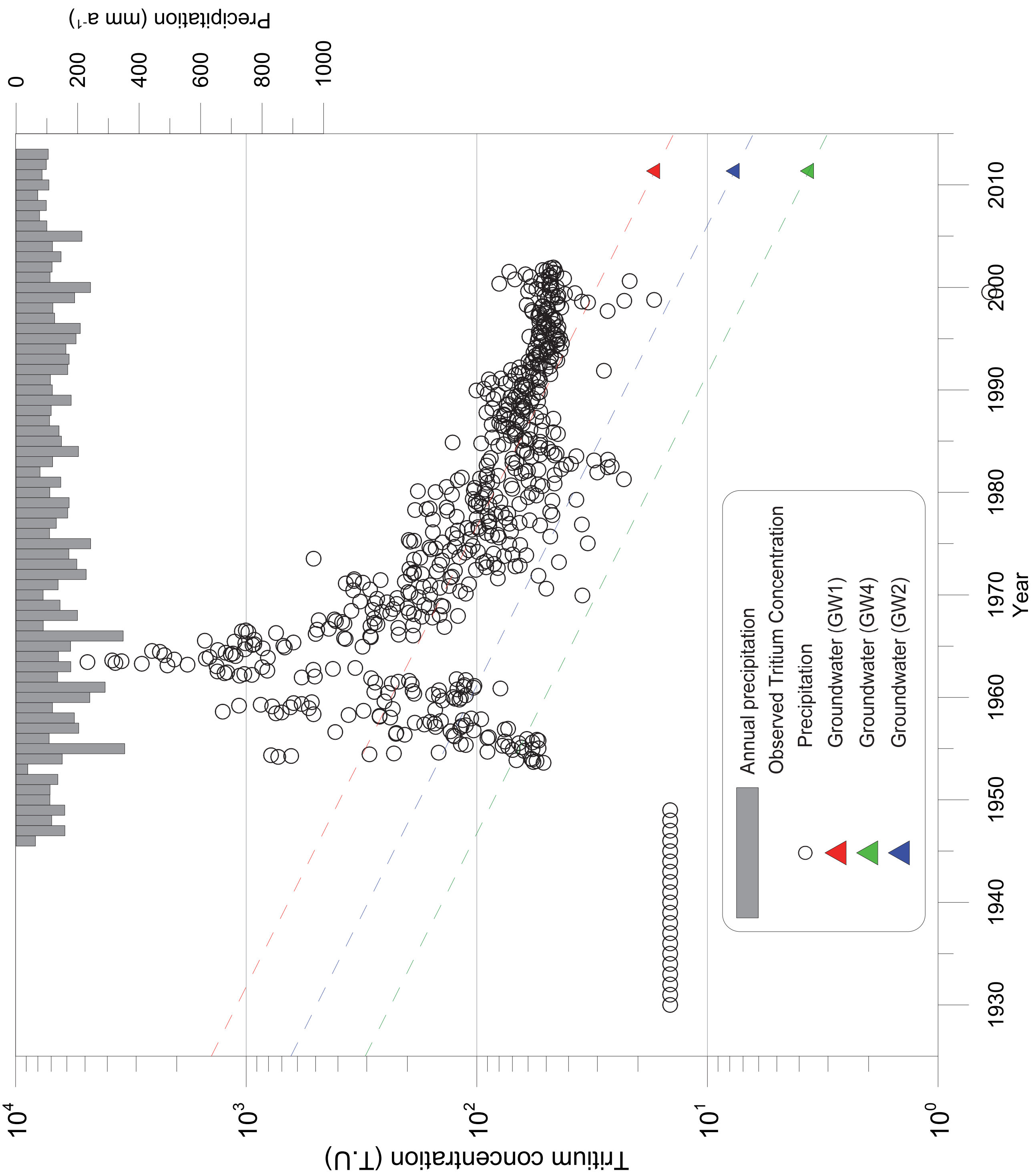
GWL (m)
T_{soil} (°C)
SWC (%)
ΣE, ΣP, Σ(P-ΔS) (mm)
Rainfall (mm day⁻¹)
Rainfall (mm 10 min⁻¹)
Date
7/1/08
1/1/09
7/1/09
1/1/10
7/1/10
1/1/11
7/1/11
ΣE
ΣP
Σ(P-ΔS)
SWC (5 cm)
SWC (10 cm)
SWC (20 cm)
SWC (30 cm)
SWC (70 cm)
SWC (110 cm)
T_{soil} (10 cm)
GW1
GW2
GW4
Rainfall

Fig. 3



Well Number
50,000
40,000
30,000
20,000
10,000
0
Livestock Number
250,000
200,000
150,000
100,000
50,000
0
1980
1990
2000
2010
Year
Camel
Horse
Cattle
Sheep
Goat
Wells in the pasture

Fig. 4


Precipitation (mm a⁻¹)
0
200
400
600
800
1000
Tritium concentration (T.U)
10^4
10^3
10^2
10^1
10^0
1930
1940
1950
1960
1970
1980
1990
2000
2010
Year
Annual precipitation
Observed Tritium Concentration
Precipitation
Groundwater (GW1)
Groundwater (GW4)
Groundwater (GW2)